THE GEOLOGY OF THE NORTHERN ALGOA BASIN, PORT ELIZABETH

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ABSTRACT

The chief object of this study was to gain insight into the geo= logical events which were connected with the earliest development of the Northern Algoa Basin. Of particular interest was the nature, origin, and stratigraphic significance of the Suurberg basalts and tuffs which are exposed along the periphery of the basin.

The main conclusions of the study are the following:

i) A marked angular unconformity separates the Mesozoic rocks (including the Suurberg volcanics) within the Algoa Basin from the underlying Paleozoic "floor". No evidence was found to support the existence of the post-Enon "Zuurberg Fault" which had been postulated by earlier workers.

ii) The Suurberg volcanic rocks antedate the overlying Uiten= hage beds, but the time interval between the two rock groups is small.

iii) The Suurberg Group can be subdivided into three rock units. The oldest, the Slagboom Formation, consists mainly of quartzite breccia, which appears to represent the initial vent-opening phase of the volcanism. This is followed by the Coerney Formation which is composed of rhyolitic ash-fall and ash-flow tuffs. The uppermost Mimosa Formation embraces tholeiitic basalt, a few dolerite dykes, and thin interbedded acidic tuffs.

iv) The volcanism was a local phenomenon within the Cape Folded Belt and took place subaerially. Eruptions were of a linear type, with the ascend of magmas probably related to pre-Uiten= hage faulting. The distribution of volcanic materials was largely controlled by the paleotopography.

v) The volcanic rocks originated from two unrelated (though contemporaneously erupted) magmas. The basaltic magma was probably generated in the upper mantle through partial mel= ting and fractionation, while an anatectic origin is indicated for the rhyolitic magma. No intermediate types occur.

vi) The Northern Algoa Basin was initiated by downwarping and crustal subsidence, which only followed after the extrusion of the Suurberg basalts.

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I INTRODUCTION

A. STATEMENT OF PROBLEM

The latest geological map of South Africa (1970) shows rocks of Cretaceous age occurring in a number of small depositional basins along the southern and eastern Cape coastal area and the adjacent interior.

The largest of these depositories, known as the Algoa Basin, covers about 4 000 km^2 in the divisions of Port Elizabeth, Uitenhage, Alexandria, Kirkwood, and Steytlerville. A special feature of this basin is the presence of volcanic rocks along its northern and northwestern periphery. The origin and age of the basalts and tuffs are at present still a matter of conjecture although published geological maps of the area tentatively correlate these rocks with the Stormberg Series of the Karoo System. The bulk of the sediments in the Algoa Basin belongs to the Uitenhage Group (or Series) which is subdivided, from bottom to top, into the Enon Formation, the "Wood Beds" and the Sundays River Formation (Truswell, 1967, p. 111).

As a result of oil exploration activities the Algoa Basin has attracted considerable attention and effort since about 1966. Geological field work, geophysical surveys, and drilling provide information which has made a fair reconstruction of the depositional history and structural evolution of this basin feasible. There are, however, certain aspects which could be partially of wholly clarified only by detailed mapping of the outer portion of the Algoa Basin. These include:

i) The nature and distribution of the volcanic rocks which crop out at certain localities along the edge of the basin;

ii) the stratigraphic relationship between the Uitenhage Group and the above-mentioned volcanic rocks;

iii) the nature of the contact between the older Cape and Karoo rocks on the one hand and the younger rocks within the basin on the other hand.

The present study attempts to fill this gap.

B. PREVIOUS WORK

It would appear from the literature that the periphery of the northern Algoa Basin has so far only received superficial field examination. Nevertheless, the volcanics which occur here and the postulated "Zuurberg Fault" have been used to support various statements regarding the regional geologic evolution of South Africa. These will be treated further on.

The first geologists to work in this part of the country, amongst others A.G. Bain, W.G. Atherstone, and R. Pinchin, are not known to have mentioned the eruptive rocks specifically. However, Atherstone and Bain were probably the first geologists to have seen them although their true nature was not appreciated at the time. In 1857 Atherstone wrote (cited by Rogers, 1906, p. 42) that the "claystone porphyry" at Mimosa was vesicular, evidently in reference to the basalts on this farm. This exposure is of historical interest, having been regarded as part of the Dwyka by Bain and Atherstone who were thus led to assign an igneous origin to the tillite (Hatch and Corstorphine, 1909, p. 300).

The first definite reference to the volcanic rocks occurring south of the Suurberg Range is a paper by Rogers (1905) in which he describes the amygdaloidal basalts, tuffs, and breccia which form outcrops along a narrow strip in the Uitenhage and Alexandria divisions. Further information, especially concerning their petrographical characteristics, was published in 1906 by the Geological Commission of the Cape of Good Hope. In both papers Rogers came to the conclusion that the volcanic rocks were of post-Uitenhage age and "....rose along the line of faulting during or after the production of the fault", referring to the "Zuurberg Fault" which he postulated at the time. Rogers (1906, p. 41) added: "...... this explanation does not account for certain facts, the presence of a band of pipe-amygdaloid in the Mimosa valley where the 'pipes' are arranged perpendicularly to a plane which dips about 20^{o} towards S.10^{o}W., and the existence of a band of scoriaceous lava parallel to the pipe rock".

In the period 1922 to 1923 the area south of the Klein Winterhoek and Suurberg ranges was geologically surveyed by S.H. Haughton and revisited by Rogers. As a result Haughton and Rogers (1924, p. 235) claimed that they had "..... found in several places conclusive evidence of the pre-Uitenhage age of the volcanic rocks, which are unconformably overlain by Uitenhage beds. There is no reason to think that the volcanic rocks rose along a fault, for it is evident that their association with faults is due to their being the lowest horizon exposed on the downthrow sides of the faults at the several places where they appeared at the surface." The same authors continued: "The new evidence ... suggests that the volcanics and the sediments immediately underlying them belong to the Stormberg series of the Karoo system, but up to the present time that has not been proved" (op. cit., p. 237). The conclusion concerning the possible age of these rocks was based mainly upon the presence of basalt and tuff fragments in overlying Cretaceous beds, the petrological similarity of the basalts with those from the Drakensberg region, and the presence of (reptilian?) bone fragments in the sediments below the basalts.

The postulated "Zuurberg Fault" and the volcanics of this area assumed further importance. The former, regarded as one of several similar post-Enon faults, was linked with the final stages of breaking up of Gondwanaland (Du Toit, 1954, p. 572). Further it was pointed out that the presence of the relatively undisturbed volcanics within the Cape Fold Belt would indicate that the main phase of folding had already been completed before the close of Stormberg times, i.e. Late Triassic to Early Jurassic (Haughton, 1935, p. 10). Also, it would appear as if the Drakensberg lava sheet extended so far southwards that the Suurberg basalt re= presented a downfaulted remnant of this once vast sheet of flood basalts (Du Toit, 1954, p. 304).

Meyer (1965) mapped a small portion of the north-eastern edge of the Algoa Ba= sin and drew attention to unusual overfolding towards the south within the Witteberg beds. Despite lack of field evidence Meyer (1965, p. 50) accepted the existence of the Zuurberg Fault and deduced that this fault had already initiated in pre-Enon times and had been subjected to renewed movement probably until shortly before the deposition of the Wood Beds (op. cit. p. 59).

In that same year the volcanics received more specific attention when J.A.H. Marais of the Geological Survey mapped about 80 km^2 south of the Suurberg Range on aerial photographs. This work was not completed but Marais provisionally came to the conclusion that the volcanics were interbedded with Enon conglomerate and that they were therefore of (Early) Cretaceous age (Annals of the Geological Survey, 1965, p. 10). It is clear that Marais correlated the quartzite breccia and conglomerates underlying the tuffs (see later) with the Enon Conglomerate.

D. Rigassi (1968), in an unpublished report on the Algoa Basin, assigned a Late Cretaceous age to the volcanics. In this connection he mentioned a potassium-argon dating on one sample as indicating an age of 80 to 90 million years, and also that thin tuffaceous layers occurred within the Sundays River beds. He further commented on the generally obscured contacts of the volcanics with both the older Paleozoic rocks and the Cretaceous beds. For this reason it was not known whether the said contacts were stratigraphical, intrusive or tectonical (op. cit., p. 7).

During 1968 the Suurberg volcanic formations were further investigated by the Geological Survey. In a field report H. Palloks came to the conclusion that the volcanics were older than the Enon, " ... but younger than the underlying conglo= merate, the age of which could not be determined". Palloks also stated that he could find no evidence for the post-Enon faulting in this area as postulated by earlier investigators.

C. PRESENT INVESTIGATION

It is evident from the foregoing résumé that various workers have held quite divergent views on the mutual relationships of the rocks along the margin of the Algoa Basin. To a large extent this may be attributed to the scarcity of good exposures and the fact that the contacts are largely obscured by superficial depo= sits and dense vegetation. For this reason it was decided right at the outset of the present investigation to cover the critical strip between the older Paleo= zoic rocks and the Enon Formation as systematically as was practically possible.

An area of approximately 1100 km^2 was mapped on aerial photographs with scales varying between 1:20 000 and 1:50 000. Certain important outcrops were mapped on even larger scales using plane table and/or enlarged aerial photographs.

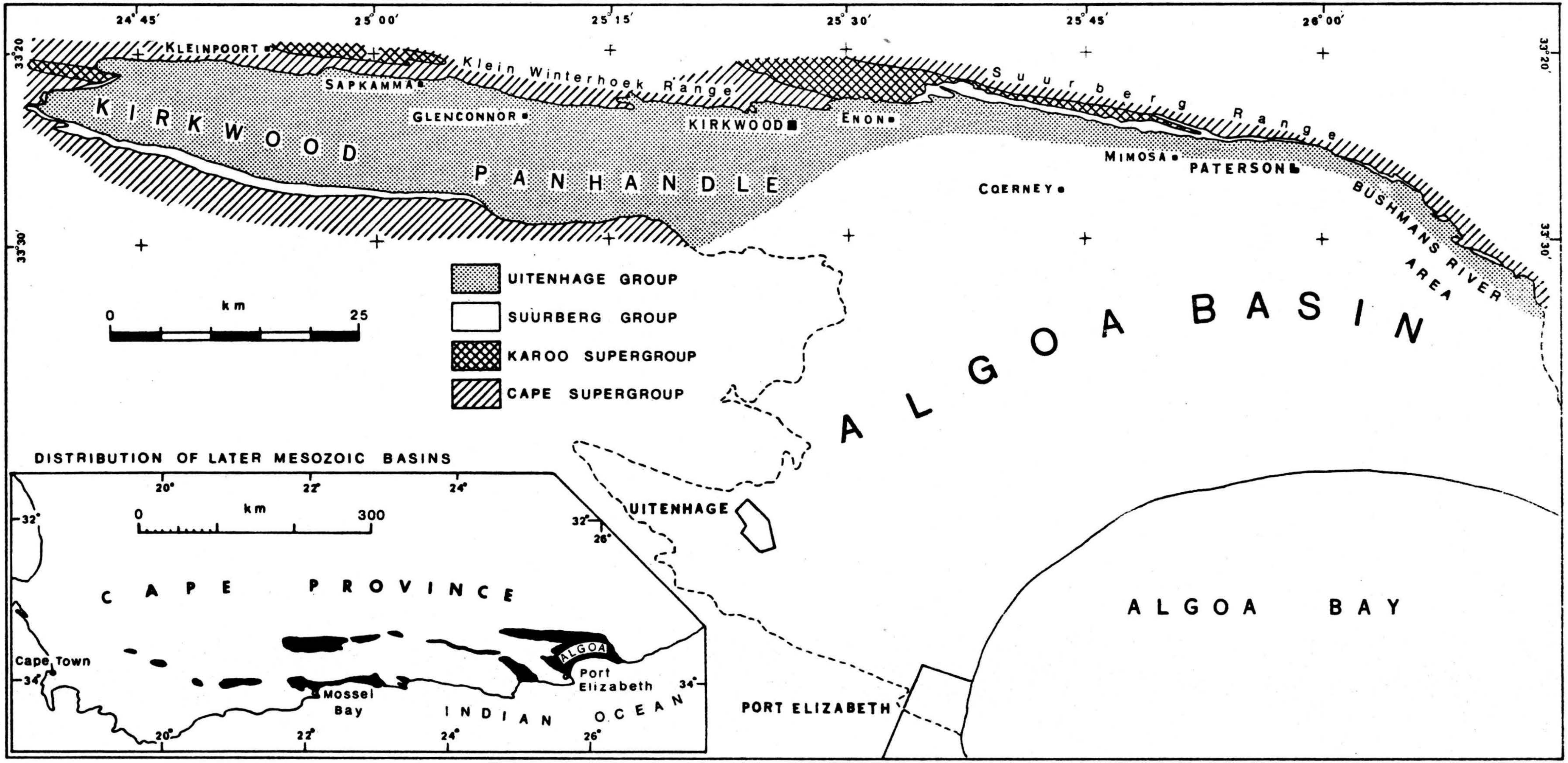


FIGURE 1. LOCALITY MAP

For convenience the mapped area is divided into three geographical portions (see Fig. 1):

i) The Kirkwood Panhandle stretching from Kirkwood westwards;

ii) the Kirkwood - Paterson area which lies between these two towns along the margin of the Algoa Basin and

iii) the Bushmans River area east of Paterson.

The geology of the two last named areas (approximately 200 km^2) was transferred to the available 1 : 18 000 sheets of the Trigonometrical Survey while their 1 : 50 000 topographical survey sheets were used as base maps for the remaining area. From these sheets most of the final maps have been compiled.

The locality names used in this paper are taken from the 1 : 250 000 topocadastral map and include names of villages and railway sidings, as well as original farm names. These names are followed by a letter in parentheses which refers to a particular column on the accompanying 1 : 100 000 geological map (Map 1).

The time spent in the field adds up to about four months, the work being done during 1970 in the course of normal duties for the Geological Survey.

It is not the intention of this work to give an exhaustive description of the geology of the investigated area. Attention will be drawn mainly to aspects with direct bearing on the problems stated in the introductory paragraphs.

D. PHYSIOGRAPHY

The area lies between 200 and 750 metres above sea level. The transition from Cape and Karoo rocks to the younger units in the Algoa Basin is often characterized by a rapid decline in height, particularly when Witteberg quartzites form the outer rim of the basin. Perhaps, in the past, the postulated post-Enon faults have been held at least partly responsible for this phenomenon. However, the variability in resistance to weathering of the different rock types appears to offer a more likely explanation.

The Suurberg Plateau is a striking physiographic feature immediately north of the Algoa Basin. This old land surface slopes gently south-eastwards and represents the product of denudation which has probably already started in the Mesozoic era.

The area is drained mainly towards the south and east by the Sundays, Groot and Bushmans Rivers and their tributaries.

From Kirkwood westwards the vegetation can be described as karooid and rock outcrops are quite accessible. Eastwards, however, the terrain is covered by dense bush and shrub which in places become virtually impenetrable so that contacts can sometimes be located only with great difficulty.

II FIELD GEOLOGY

The stratigraphic subdivision used in this study is shown in Table 1.

TABLE 1

STRATIGRAPHIC SUBDIVISION*

ERA	SUPER-GROUP	GROUP	FORMATION	LITHOLOGY (Main types only)
CAINOZOIC				Alluvium, blown sand; terrace gravels; conglomerate, sandstone, marl (Alexandria Beds)
MESOZOIC		UITENHAGE	"Wood Beds"	Mudstone, sandstone
			Enon	Conglomerate, sandstone (including the Basal Enon Sandstone)
		SUURBERG	Mimosa	Basalt, dolerite, tuff
			Coerney	Tuff
			Slagboom	Breccia, conglomerate
PALEOZOIC	KAROO	ECCA		Shale, greywacke
		DWYKA		Tillite
	CAPE	WITTEBERG	"Upper Shales"	Shale, sandstone
			"Main Quartzites"	Orthoquartzite, sandstone (including the "White Streak")
			"Lower Shales"	Siltstone, shale, sand=stone, orthoquartzite (including the Driekuilen Sandstone)
		BOKKEVELD		Siltstone, shale, sand=stone

*The South African code of stratigraphic terminology and nomenclature (Trans. geol. Soc. S.Afr., 74, 111-131) became available only after the present thesis had been virtually completed. It is realized now that, according to the above-men=tioned code, the term "Basal Enon Sandstone" is technically not correct. The same applies to the subdivisions of the Witteberg Group. The author refrained from introducing new formational names, except in the case of the Suurberg Group which had been investigated in considerable detail.

A. THE PALEOZOIC ROCKS

The Paleozoic is represented by rocks belonging to the Cape and Karoo Supergroups. They form the outer margin of the Algoa Basin and can be regarded as the "basement" which underlies the younger units occurring within the basin.

Rocks of the Bokkeveld, Witteberg, Dwyka, and Ecca Groups are exposed in the mapped area. Their lithology and stratigraphy have been treated in some detail by Johnson (1966), Venter (1969), and others, and therefore only points relevant to the present study will be briefly noted.

1. The Bokkeveld Group

Johnson (1966, p. 15-22) and Theron (1970, p.203) subdivided the upper part of the Bokkeveld east of Steytlerville as shown in Table 2.

TABLE 2

SUBDIVISION OF THE BOKKEVELD GROUP (UPPER PORTION)

Conventional	Theron (1970)	Johnson (1966)	
(Witteberg)	(Witteberg)	(Witteberg)	
?	Vondeling Fm.	Erekroons Fm.	
?	Driekuilen Fm.	Driekuilen Fm.	
Fourth Shale	Adolphskraal Fm.	Adolphskraal Fm.	Red Shale Member Silt-shale Member Argillaceous Shale Member

Further work convinced Johnson that the Red Shale Member and overlying formations had closer lithological affinities to Witteberg rocks and consequently he decided that the top of his Silt-shale Member should rather be taken as the top of the Bokkeveld Group (personal communication). This suggestion agrees with conclusions reached by Soekor geologists (Venter, 1969, p. 7) but not with Theron (1970) who regards his Bokkeveld-Witteberg boundary as "... a compromise between convention, facies variations and petrography" (op. cit., p. 197).

However, for the present investigation it is only of some importance to have a mappable lithological boundary south of the Kirkwood Panhandle in order to bear out the structure. For this purpose the base of Johnson's Red Shale Member was mapped as the base of the Witteberg Group. Consequently a portion which is shown as Bokkeveld on existing maps of the area is now tentatively included with the lowermost Witteberg.

2. The Witteberg Group

Conventionally the Witteberg of the southern Cape Province is subdivided into the "Lower Witteberg", the "Main Witteberg", and the "Upper Witteberg" (Theron, 1962, p. 364).

a) Lower Witteberg

As explained above the Lower Witteberg starts with Johnson's Red Shale Member because white quartzite bands and red micaceous siltstones, which are not usual for the Bokkeveld, make their appearance in this unit. In addition the Lower Witteberg consists of greyish and reddish well-bedded silty shales and brownish to greyish sandstones and orthoquartzites. Plant imprints, spirophyton, and other signs of biologic activity are typical of many bedding planes.

The Driekuilen Sandstone (Formation) is found within the Lower Witteberg and consists of prominent sandstones and quartzites some 100 metres thick. It is shown on the accompanying maps as a marker horizon which occurs approximately 500 metres below the top of the Lower Witteberg. This latter unit has a total thickness of about 975 metres, measured by Johnson (1966) near the western extremity of the mapped area.

The Lower Witteberg is highly folded and forms the outer boundary of the Kirkwood Panhandle in the south and west.

b) Main Witteberg

This unit consists mainly of brownish and white orthoquartzites and sandstones. The uppermost 30 metres is composed of a conspicuous white thick-bedded orthoquartzite, the "White Streak", which marks a sharp boundary with the overlying "Upper Witteberg Shales". From Paterson eastwards the White Streak becomes more or less detached from the main quartzite mass through intervening shales (Haughton, 1928, p. 12).

The thickness of the Main Witteberg progressively increases from west to east along the margin of the Algoa Basin. At Losberg, on the farm Drie Kuilen (A), it amounts to 275 metres (Johnson, 1966, p. 26); at Uiepoort, two kilometres northeast of Kirkwood, 580 metres; in the vicinity of Grahamstown it is esti= mated at about 700 metres (Mountain, 1946, p. 11).

The Main Witteberg exhibits intense folding, overfolding, faulting and thrusting caused by the Cape Orogeny (De Villiers, 1944).

c) Upper Witteberg

For the purpose of this study all rocks occurring between the White Streak and the base of the Dwyka Tillite are regarded as Upper Witteberg. Loock (1967) has proposed a twofold subdivision of this unit into the Lake Mentz and Kommadagga Formations and shows that they reflect the transition from Krynine's peneplanation stage to a geosynclinal stage.

Dark greyish to greenish shales and siltstones with subordinate sandstone and orthoquartzite are typical of this unit. In the Kirkwood-Paterson area the uppermost sediments are usually greywackes which pass into tillite higher up.

The Witteberg-Dwyka boundary northeast of Kirkwood in the vicinity of the Uye River appears rather irregular in the field. Some pre-Dwyka erosion might have taken place locally.

The thickness of the Upper Witteberg is variable but most values fall within the range 300 to 600 metres.

This unit appears to have been subjected to the same tectonic deformation as the Main Witteberg.

3. The Dwyka Group

It appears desirable, on lithostratigraphic grounds, to include the "Upper Dwyka Shales" of previous authors with the Ecca Group (Winter and Venter, 1970, p.396). This would also simplify mapping in the southeastern Cape where the "White Band", which marks the top of the former unit, does not show up as prominently as elsewhere.

For the purpose of this study the Dwyka consists of tillite only with a few thin interbedded shales developed locally, especially towards its base. As defined above the Dwyka should be given formation rank, but to avoid confusion the designation "Dwyka Group" is provisionally retained.

Lithologically the tillite of this area agrees with the descriptions of other authors (e.g. Haughton, 1935, p. 15). The thickness of the Dwyka Tillite remains close to 600 metres everywhere in the surveyed area. Due to its massive character, folding does not show up so well in the tillite but closer investigation confirms the presence in it of steep folding. The Dwyka in this area is often a more prominent topographical feature than the underlying Upper Witteberg which appears to be more susceptible to weathering. This fact is also borne out by the distribution of the Mesozoic rocks at De Vlei (A) and north of Mimosa (F).

4. The Ecca Group

In the Kirkwood-Paterson area the outer margin of the Algoa Basin is partly formed by Ecca greywackes and shales which are not indicated on existing maps. A stratigraphic thickness of at least 500 metres is present north of Enon Mission Station. Of this figure the lowermost 150 metres could be ascribed to the "Upper Dwyka Shales" of previous authors, while the remaining 350 metres, or more, belongs to the "Lower Ecca" or Ecca Pass Formation of Johnson (1966).

The greywackes and dark grey shales display the same features as the basal rocks of the Ecca elsewhere in the southern Cape. They are therefore geosynclinal deposits formed as a result of turbidity currents (Truswell and Ryan, 1969).

The location of this Ecca is rather interesting because it occurs enclosed within the Cape Folded Belt and not just immediately north thereof as is other=wise the case. This fact confirms the idea that the boundary of the Southern Karoo Geosyncline extended quite far southwards and also proves that the major events of the Cape Orogeny postdate Lower Ecca times. The Cape Folding can thus be regarded as a manifestation of an orogenic stage which followed upon the Dwyka-Ecca geosynclinal stage as visualized in Krynine's tectonic cycle.

B. THE SUURBERG GROUP

This is the name proposed for a sequence of mainly pyroclastics and basalts which occur south of the Klein Winterhoek-Suurberg Mountain range. Outcrops of these rocks are restricted to the periphery of the Algoa Basin and, as far as known, similar rocks do not occur elsewhere in the south-eastern Cape Province. The Suurberg Group overlies rocks of the Cape and Karoo sequence discordantly and is in turn overlain by the Uitenhage Group. A threefold lithostratigraphic subdivision of the Suurberg Group is proposed in this thesis(cf. Table 1).

In the following discussion the petrography of the finer grained rocks of this group is held over until Chapter III, while major structural features will be treated in Chapter IV. For an explanation of the nomenclature for pyroclastics used below the reader is referred to the Appendix.

1. The Slagboom Formation

a) Definition

The name Slagboom Formation is proposed for a unit, composed mainly of breccia and conglomerate, which forms the lowermost part of the Suurberg Group. Rocks of this formation are best exposed on the farm Slag Boom (E) where the unconformable lower contact with Paleozoic rocks can also be seen. The upper limit, which is mostly gradational and rather poorly exposed, is defined as the line where the rudites lose their predominance to tuffs of the overlying Coerney Formation.

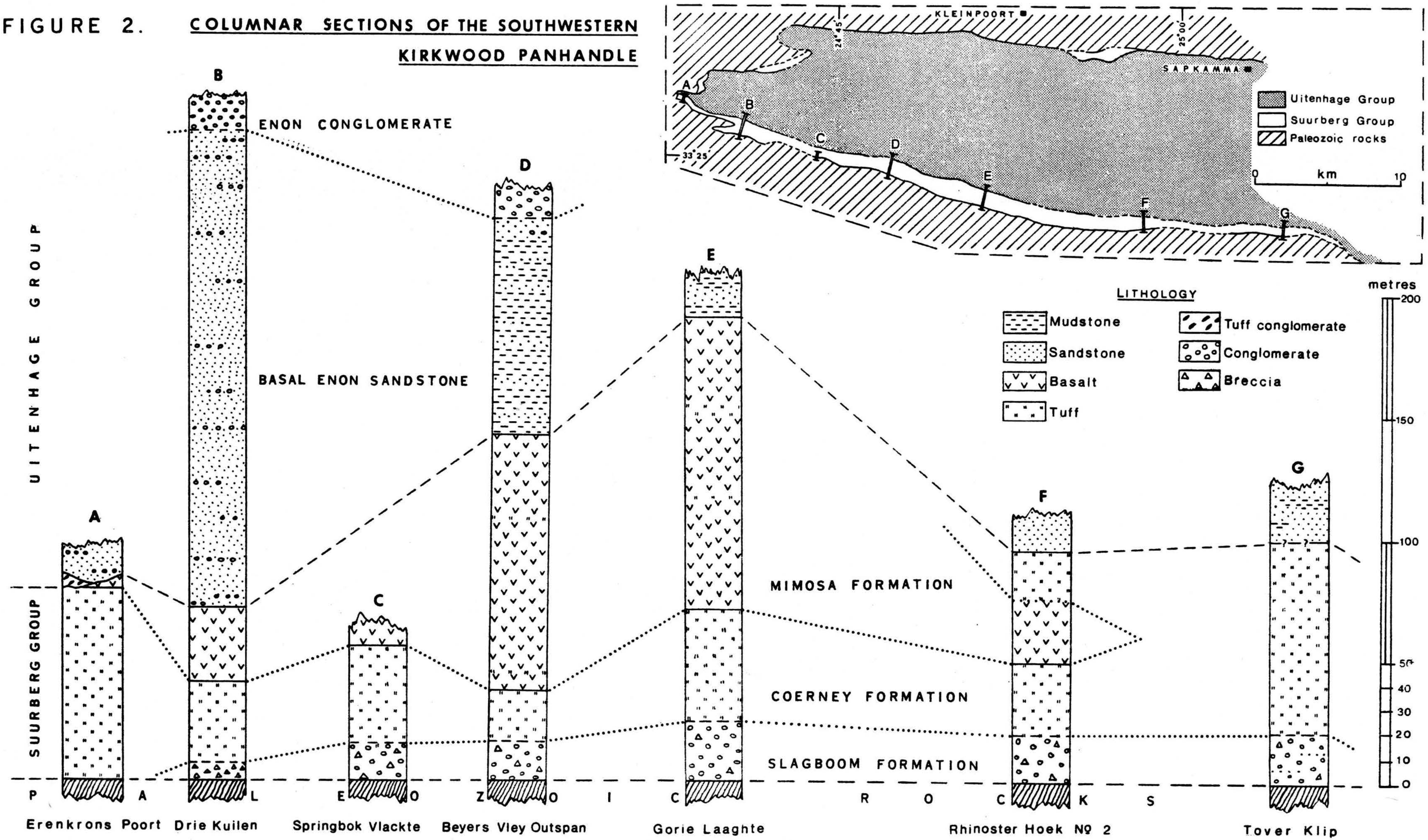
b) Distribution and Lithology

The Slagboom Formation forms virtually continuous, though poor outcrops along two narrow strips, namely, between Tover Klip (C) and Drie Kuilen (A) in the southern Panhandle area, and between Enon Mission Station (E) and Zand Vlackte (F) in the Kirkwood-Paterson area (see Map 1).

In the southern Panhandle outcrop area conglomerate predominates while the same is true of breccia in the Kirkwood-Paterson area.

The conglomerate consists of closely packed, poorly sorted pebbles and cobbles ranging in size from 2mm to about 25cm, pebbles of 0,5 to 5 cm being the most abundant. Clasts of light grey orthoquartzite, greenish grey micaceous siltstone, white vein quartz, chalcedony, and, much rarer, tuff fragments occur.

FIGURE 2. COLUMNAR SECTIONS OF THE SOUTHWESTERN KIRKWOOD PANHANDLE



Most pebbles are moderately rounded with rather low sphericity. Discoidal and blade shapes (Zingg's classification) are common. The matrix material is sandy to argillaceous and may also be tuffaceous in part. Secondary silica and red iron oxides are the cementing agents.

The breccia differs from the conglomerate mainly in the higher degree of angularity displayed by the clasts. In addition, in the Kirkwood-Paterson area it usually contains relatively more tuff and a smaller variety of rock types. In this latter area light grey quartzite occurs as principal and, sometimes, only framework constituent of the breccia. In this case the matrix is composed of the same, but much finer, quartzite particles.

Extensive search revealed only one rounded tillite fragment within the breccia at Slag Boom (E). Otherwise tillite is conspicuously absent from the conglomerates and breccia. Except for some tuff fragments and the tillite pebble mentioned above the clasts appear to represent exclusively rock types of the Cape Supergroup, and more particularly the Witteberg Group.

Locally lenses or troughs of better rounded pebbles occur within the breccia. Conversely, some very angular fragments occur within the conglomerate of the southern Panhandle area and towards the west, on the farm Drie Kuilen (A), only breccia was seen (Plate 1 A).

Vertical to near-vertical parallel cracks which pass through matrix and clasts alike are a conspicuous feature of the Slagboom rudites which will be discussed further in Chapter IV.

Good outcrops of the Slagboom Formation are rare. The conglomerate is best seen on the farms Springbok Vlackte No. 6 (AB), Tover Klip (C), and Slag Boom (E), while the breccia is best exposed on Drie Kuilen (A), Enon Mission Station (E), and Slag Boom (E). At other localities the rocks of the Slagboom Formation are mostly weathered to loose gravel.

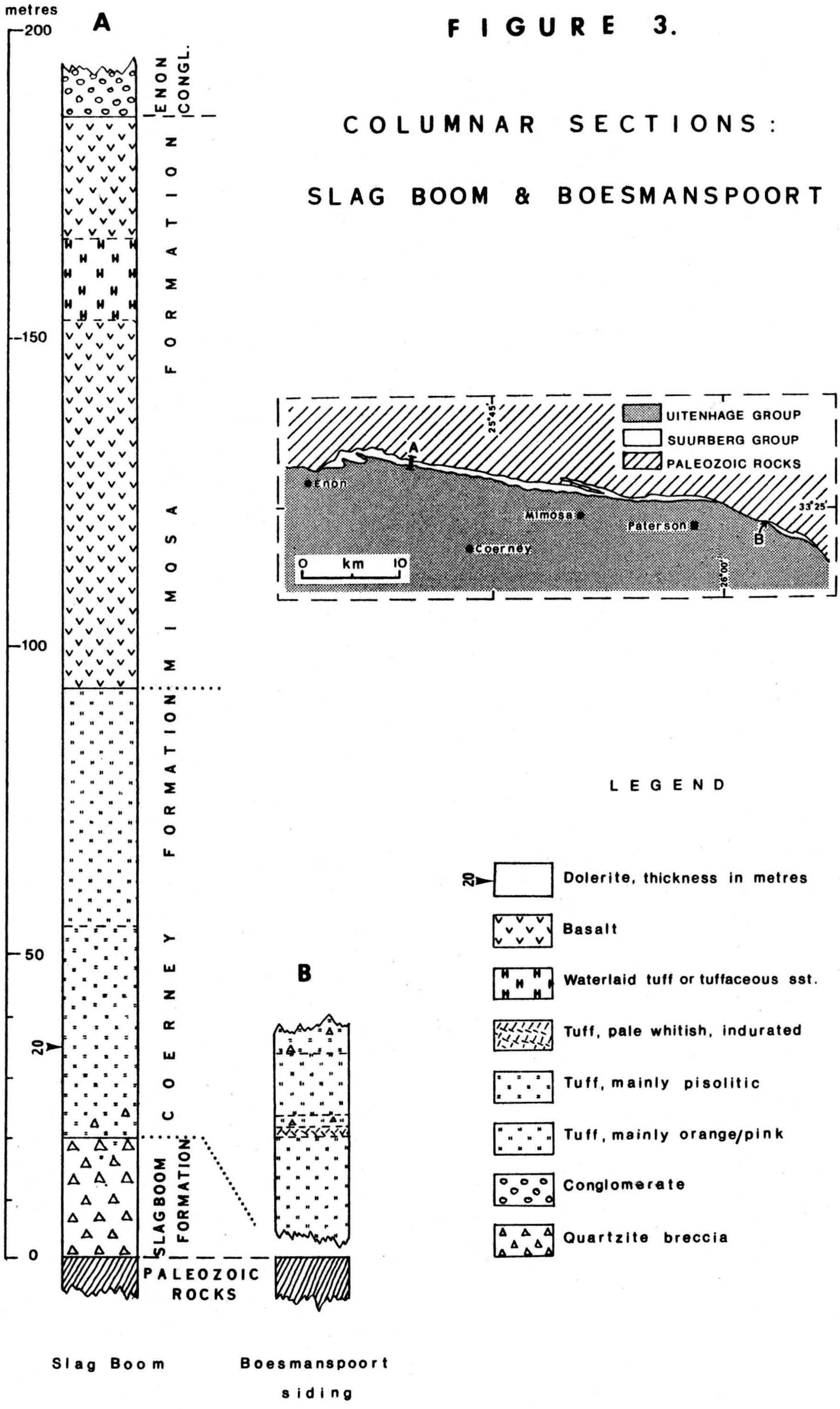
The breccia appears to be largely unbedded. Poor bedding can sometimes be seen in the conglomerate, particularly where thin sandstone bands occur, as for instance on Tover Klip (C).

The stratigraphic thickness of the breccia, calculated from two boreholes on the farm Unamore (F), varies between 20 and 30 metres. Farther westward on Slag Boom (E) the thickness appears to be of the same order. A reliable thickness for the conglomerate in the southern Kirkwood Panhandle is difficult to determine. Indications are that it would not exceed 30 metres and perhaps a figure of between 15 and 20 metres would be more realistic in most cases (see also Figures 2 and 3).

The Slagboom conglomerates in many ways resemble those of the Enon Formation, which will be described later. However, the former tend to have smaller clasts and more abundant tuff material incorporated.

FIGURE 3.

COLUMNAR SECTIONS: SLAG BOOM & BOESMANSPOORT



c) Probable origin

The breccia is obviously not the result of normal fluviatile processes and therefore three other possible modes of origin were considered in the field:

i) The product of pre-volcanic folding and faulting, i.e. cataclastic breccia (Pettijohn, 1957, p. 281);

ii) the product of weathering with or without gravity sliding, i.e. talus breccia;

iii) the result of volcanic explosions, i.e. explosion or pyroclastic breccia.

Ice and mud flows were never seriously considered as possible mechanisms of origin because of the low matrix : framework ratio and the oligomictic nature of the breccia.

Along most of the Kirkwood-Paterson outcrop area the breccia is underlain by Dwyka tillite. In spite of this fact practically no fragments of tillite have been noted in the breccia so that a cataclastic origin is ruled out.

It is more difficult to decide between the two remaining possibilities; even a combination of the two cannot be excluded.

A talus origin for the breccia warrants serious consideration because, as will be shown later, the area now covered by the Mesozoic rocks was one of mountainous relief (see also Plate 3 F). Earth tremors before and during the volcanic eruptions could have caused the accumulation of slide breccia in low lying areas followed by flows and falls of volcanic materials. Local reworking by fluvial agents would account for the conglomerate within the Slagboom Formation.

On the other hand there is evidence which would suggest a pyroclastic origin for the breccia:

i) This formation occurs at the base of an indisputable volcanic sequence of rocks.

ii) Volcanism was already in progress during its deposition, as fragments of tuff within the breccia testify. It should also be noted that the upper contact with the overlying tuff unit is usually gradational at the few places where it is exposed.

iii) There is a marked parallelism in distribution between the Slagboom rudites and the Mimosa basalt (see Figs. 2 & 3), which contrasts with the wider distribution of the tuffs. The individual basalt flows probably never attained significant thicknesses, as will be shown in a following section, so that one suspects that the basalt and breccia might indicate proximity of the volcanic vent(s).
In the Bushmans River area, where no basalt occurs, the Slagboom Formation is conspicuously absent although the topographical relief

here must have been similar to that elsewhere along the margin of the basin.

iv) On Buffelskuil (F) the breccia, which is composed almost solely of angular (Witteberg) quartzite fragments, occurs 2 km south of the nearest visible quartzite source.

v) North of Mimosa (F) a ridge of Dwyka tillite, which must have existed in prevolcanic times, separates two occurrences of similar quartzite breccia (see Map 2). The southern of the two occurrences is virtually cut off from Witteberg sources to the north by this ridge. Although the breccia flanks a hill composed of Dwyka tillite no fragments of the latter have been found.

vi) Quartzite fragments, similar to those constituting the breccia, occur sparsely distributed through some tuffs higher up in the Suurberg succession. They are particularly associated with pisolitic tuffs which are believed to fall close to the volcanic source (Moore & Peck, 1962). The quartzite fragments are obviously accidental ejecta derived from the pre-volcanic basement. It would be logical to expect more of this type of ejecta towards the base of a volcanic sequence.

Apparently a likely explanation for the above features would be that the breccia was produced and dispersed as the result of volcanic explosions. Pyro= clastic breccia need have no lithological relation to the immediately underlying rock and its components can be propelled over fair distances and small topo= graphical obstacles without significant rounding. Gorshkov and Dubik (1970), for example, describe large volcanic bombs and blocks which fell up to 10 km from the centre of eruption.

Rittmann (1962, p. 45, 90) uses the term "vent-opening breccia" for deposits which result from initial eruptions or perforations and which occur typically at the base of a volcanic pile. When first of all only gas breaks through, the ejected material consists of fragments of the strata penetrated, with little or no juvenile volcanic material.

Examples of such breccias which were produced during the initial stages of volcanism are provided by Rittmann (1954), Pande and Gupta (1965), and Keller (1969), amongst others. Hughes (1960) mentions explosion breccia which occurs at the base of felsic volcanics on the Isle of Rhum along the west coast of Scotland. The clasts of this breccia were derived from the underlying country rock and are often significantly rounded. This rounding Hughes ascribes to attrition due to the continued passage of gases through rocks partially broken and cracked by initial explosions (op. cit., p. 120). Reynolds (1954) furnishes further examples of rounded breccias which appear to have been produced by streaming of particle laden gas (fluidization).

The conglomerate associated with the Slagboom breccia could therefore be accounted for either by gas attrition or by reworking in local stream channels, or possibly both. The last mentioned agent could have introduced a certain amount of external material.

A final conclusion on this matter would require additional field and laboratory investigation.

2. The Coerney Formation

a) Definition

The name Coerney Formation is proposed for the unit of tuffs and tuffaceous rocks that overlie the Slagboom Formation. Where rocks of this latter formation are not present the Coerney Formation directly overlies the older Paleozoic strata. The upper limit is taken at the base of the lowest basalt flow.

This formation is typically developed in the area north of Coerney railway siding, especially in the vicinity where the Coerney River cuts through the foothills of the Suurberg.

b) Distribution and Lithology

The Coerney Formation enjoys a wider distribution than any of the other two formations of the Suurberg Group. It is the only formation of this group that crops out in the Bushmans River area.

The outliers of tuff which are preserved in small depressions within the Klein Winterhoek range well above other rocks of the Suurberg Group on the farm Blaauwbosch Kuil (B) are particularly interesting (see Map 1). These occurrences prove the existence of a well developed mountainous relief in early Suurberg times.

The Coerney Formation consists essentially of well consolidated tuffs. Three main types, according to megascopic appearance, have been distinguished in the field. These are:

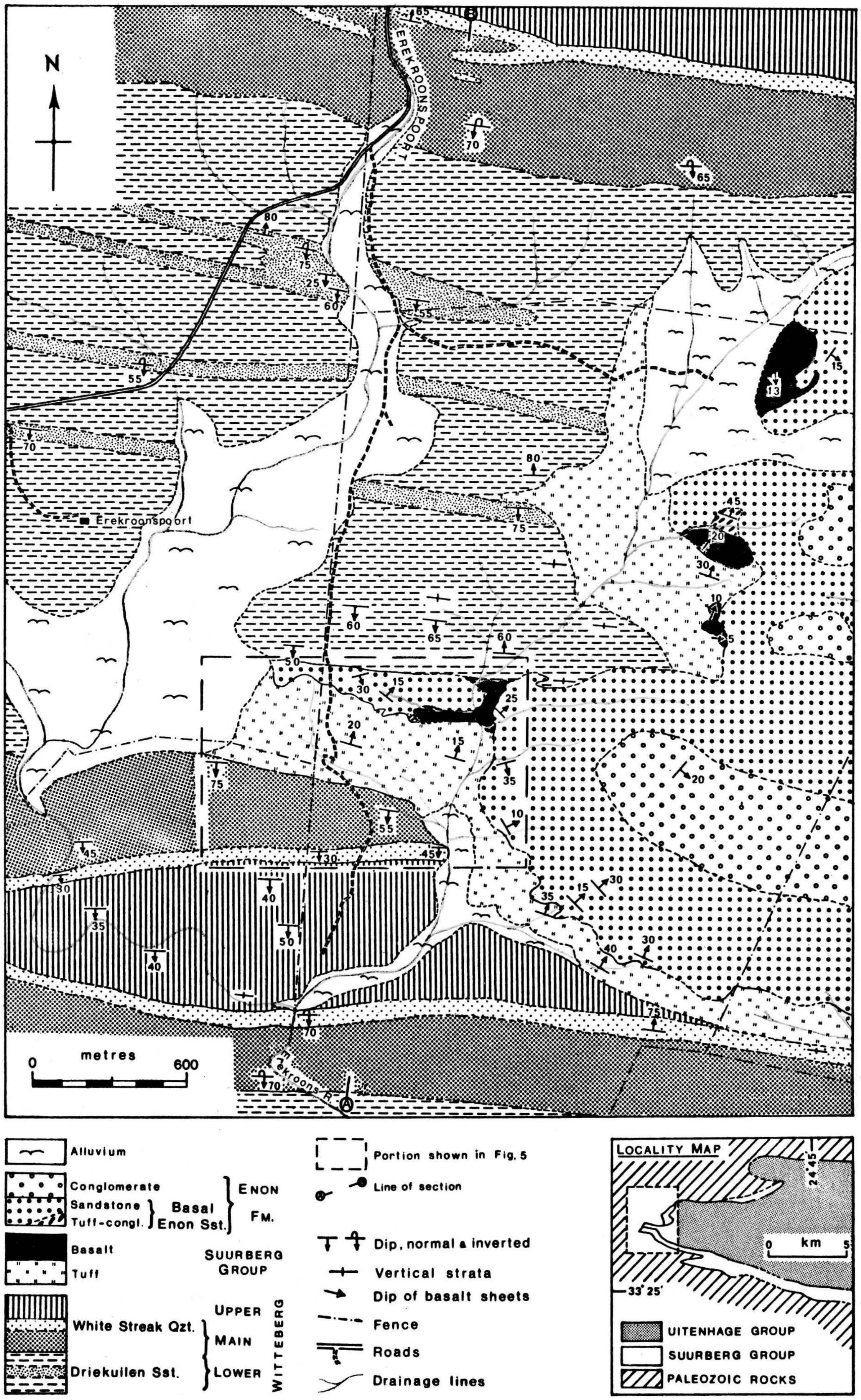
i) Pale orange or pink tuffs, which are by far the most abundant;

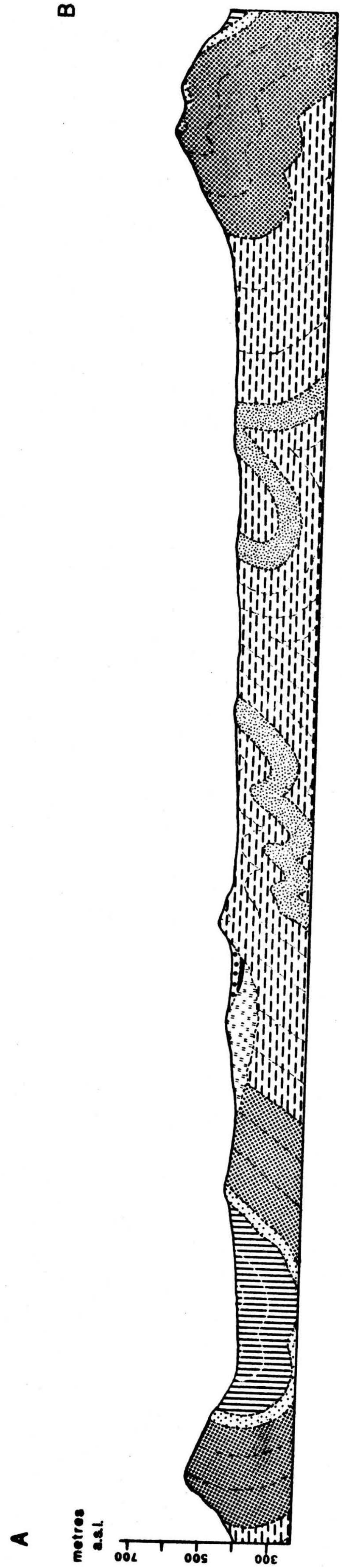
ii) pisolitic tuffs; and

iii) pale whitish tuffs.

The pale orange or pink tuffs – sometimes the colours are variegated — are always fine-grained. Small, pebble-sized fragments of quartzite and chalcedony occur sporadically in the tuff. In places where it overlies the Paleozoic rocks directly, as for instance in the Erekroons Poort area (Fig. 4), this tuff is often discoloured and cemented by red iron oxides.

FIG. 4. GEOLOGICAL MAP OF THE EREKROONSPOORT AREA





HORIZONTAL SCALE = VERTICAL SCALE

SECTION A - B

The pisolitic tuff is composed of well rounded to slightly irregular pisolites, or accretionary lapilli, with diameters generally ranging from 1 mm to 5 mm. The individual pisolites consist of fine tuff material which is sometimes arranged in concentric shells (see Plate 4B). In any given hand specimen pisolites usually fall within a fairly narrow size range. They are mostly closely packed with the void spaces wholly or partly filled with volcanic ash, chalcedony, zeolite, or, more rarely, calcite. Angular quartzite clasts are common in the pisolitic tuffs, though in small quantities only (perhaps 0 - 5% by volume). The natural rock has a rough irregular surface and is often quite porous.

Dark grey spheroidal to slightly ovaloid pellets up to 5 mm across also occur in pink tuffs on Slag Boom (E) (see Plate 1E). These structures, as well as the pisolites mentioned above, are regarded as balls of fine volcanic ash that formed in water-charged volcanic ash clouds in a manner similar to hail-stones (Moore and Peck, 1962).

The pale whitish tuffs are always extremely fine-grained. Some have small irregular patches of a soft, black soot-like substance which could not be identi= fied positively. In places this tuff is indurated to form a thin hard layer, as for instance at Boesmanspoort (G) and Woodbury (G).

Figure 5 shows the interrelationship of the various types of tuffs on Erenkrons Poort (A) where particularly good exposures occur. The tuffs show considerable variation within short distances, both laterally and vertically.

In many of the tuffs bedding is poor or virtually imperceptible, while joints or cleavage planes perpendicular to the bedding are commonly developed. Some cleavage planes were probably mistaken for true bedding in the past, with the result that faults had to be postulated locally to explain the apparently abnormal attitude of the tuff (Engelbrecht et al., 1962, p. 11; Haughton, 1935, p. 18 & 20).

Occasionally well bedded and, more rarely, cross-bedded clastics are found within the Coerney Formation. These are usually thin tuffaceous sandstones or small-pebble conglomerates which are composed of reworked pyroclastics, in addition to some externally derived material. Some are waterlaid tuffs with little or no addition from external sources. It is difficult to distinguish between all these different variations in the field. Waterlaid and reworked tuffs appear to be most common in the western portion of the Kirkwood-Paterson area but they have also been noted at several other localities. It is estimated that these tuffaceous clastics form only a small percentage, perhaps less than 10%, of the Coerney Formation.

Locally some agglomerate occurs within the Coerney Formation of the Kirk= wood-Paterson area. The agglomerate consists of unsorted angular to somewhat rounded tuff fragments of lapilli dimensions set in a matrix of volcanic ash. The tuff fragments are usually of the pale orange or pink variety. Angular

FIGURE 5

GEOLOGICAL MAP OF A PORTION OF ERENKRONS POORT

(SEE FIGURE 4)

LEGEND

Superficial deposits

Sandstone, light grey
Mudstone, red & green
Conglomerate lenses
Tuff-conglomerate
} BASAL ENON SANDSTONE

Basalt, amygdaloidal
Tuff: orange/pink
pisolitic
pale whitish
Red "contact" tuff
} SUURBERG GROUP

WHITE STREAK } UPPER
MAIN
LOWER
WITTEBERG

Contacts: well exposed, covered
Strike & dip of layers
Farm road
Fence
Water course
Form line contours (15m spacing)
469 Spot height (metres a.s.l.)

quartzite clasts are also present as a rule but in subordinate amounts.

On Slag Boom (E) and Gorie Laaghte (B) indications of fossils were found in tuffs very close to the top of the Coerney Formation (see Plate 1B). Personnel of the Bernard Price Institute in Johannesburg identified these as possible Dinosaur bone fragments (personal communication).

Borehole data from the farm Unamore (F) reveals that the Coerney Formation has a variable thickness between 35 and 50 metres. Stratigraphic thicknesses at other localities are shown in Figures 2 and 3, and are estimated to be accurate within 15%.

3. The Mimosa Formation

The name Mimosa Formation is proposed for the uppermost unit of the Suur=berg Group which comprises principally basalt with subordinate dolerite and thin interbedded tuffs. It is typically developed in the area north of Mimosa rail=way siding where it probably attracted the attention of geologists for the first time.

The lower limit of the Mimosa Formation is taken at the base of the lower=most basalt which appears to have a conformable relationship with the underlying tuffs. Basalt also forms the upper boundary of this formation with the overlying Enon sediments except at Rhinoster Hoek no. 2 (B) where tuff occurs above the uppermost basalt (see Fig. 2). The Mimosa-Enon contact will be discussed in more detail later.

For the distribution of the Mimosa Formation the reader is referred to Map 1. Its thickness at Unamore (F), determined from borehole data, is 90 metres. From this locality eastwards the basalt thins out rapidly and on Duncairn (F) inliers of Coerney tuffs are exposed. For other localities thicknesses are shown in Figs. 2 and 3.

a) Basalt

In a fresh hand specimen the rock has a greyish red (5 R 4/2)$^{+}$ to brownish grey (5 YR 4/1) colour which weathers to various shades of greenish and olive grey. The granularity varies from very fine to microcrystalline.

Vesicular, amygdaloidal, and more massive varieties of basalt occur. The vesicles are irregularly shaped; they usually measure between 1 mm and 5 mm across and are wholly or partly filled with zeolite, chalcedony, or calcite. Many vesicles are lined with a green mineral which has been identified as celadonite.

$^{+}$Rock colour symbols are given according to the Rock-color Chart issued by the Geological Society of America.

Xenoliths are very rare in the lava. On Drie Kuilen (A) and also on De Vlei (A) an inclusion of baked siltstone, evidently derived from the underlying Paleozoic rocks, was noted. On Beyers Vley Outspan (B) autoliths are fairly common in the basalt.

As far as other primary structures are concerned the basalt appears rather featureless. The tops and bottoms of flow sheets are in most cases not readily recognizable. Individual layers are perhaps several metres thick. The only indication of linear flow is slightly elongated gas holes when seen in plan at some places.

Small occurrences of scoriaceous lava were seen on Slag Boom (E) and Thornleigh (F). At the latter locality some pipe amygdales are also present. No instances of pillow or ropy lava or other similar structures have been recorded from anywhere in the area.

A particularly significant outcrop pattern north of Mimosa would suggest a duplication of the basalt and underlying pyroclastics (see Map 2). This feature is evidently not the result of folding, and a determined search provided no evidence of faulting. Of further interest is the fact that the northernmost basalt out= crops form a row of unconnected erosion relicts with no signs of a feeder dyke.

A study of Map 2 will show that the most plausible, if not only, explanation for this distribution is dependent upon the paleotopography. Roughly E.S.E. trending valleys and ridges had been sculptured in the Paleozoic rocks during pre-Suurberg times through erosion which was controlled by the under= lying lithology and structure. These valleys were subsequently partly filled by pyroclastic material which provided a fairly level surface for the succeeding lava flows. North of the present day Mimosa the lava apparently flowed round the nose of a ridge into a tributary valley which had a floor sloping slightly south-eastwards.

The present attitude and significance of this outcrop will be discussed further in Chapter IV.

b) <u>Dolerite</u>

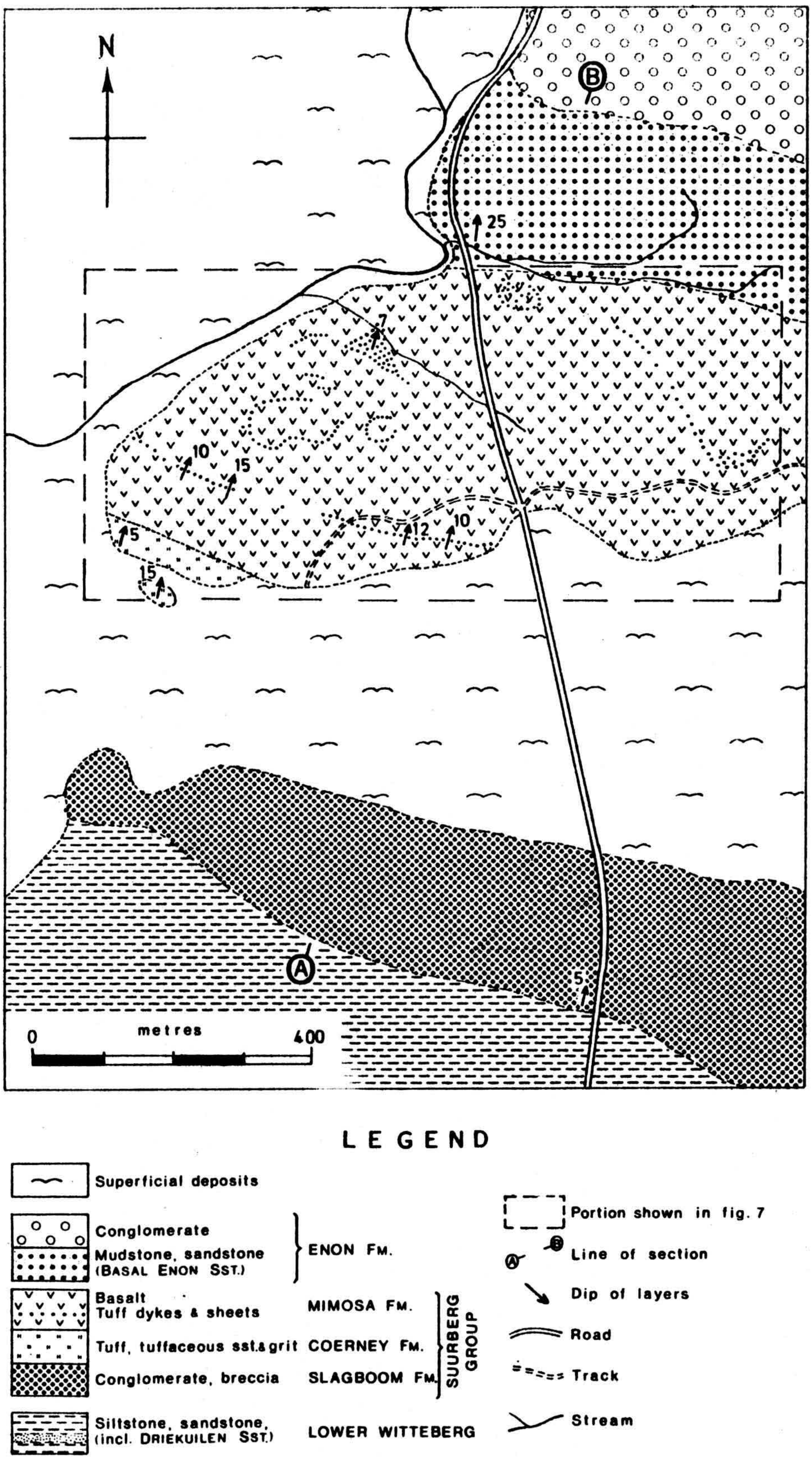
On Slag Boom (E) and also on Enon Mission Station (E) dolerite cuts through the tuff underlying the Mimosa basalt. The dolerite was distinguished from massive looking lavas only when its intrusive field relationship could be observed and for this reason other occurrences have probably been overlooked.

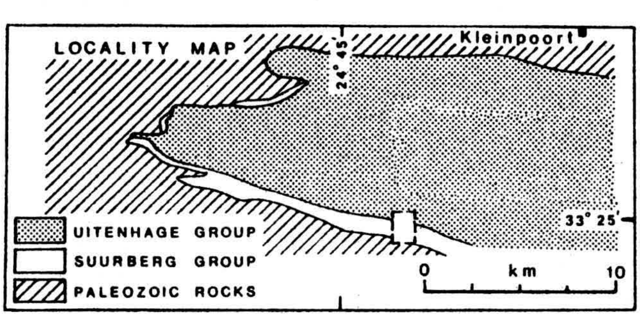
In hand specimen the rock is fine- and even-grained with an olive-black (5 Y 2/1) colour when fresh. The specific gravity is 2,85 (± 0,01). Spheroidal exfoliation as well as reddish brown decomposed outer rims surrounding fresh cores are characteristically developed.

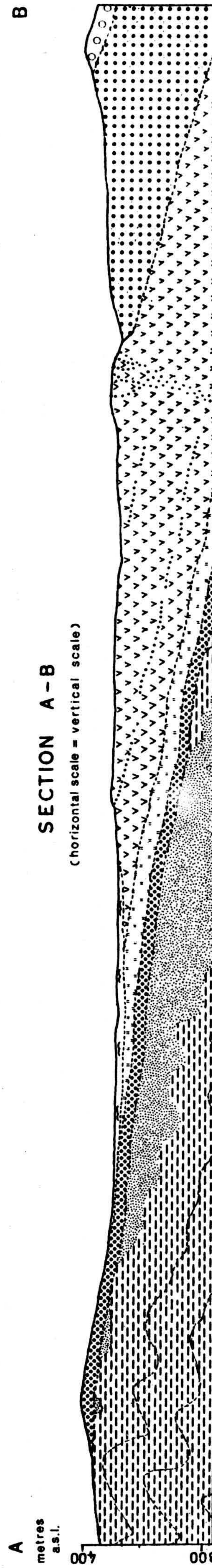
The dolerite is best exposed on Slag Boom (E) where a dyke, several metres wide, can be followed intermittently over a distance of at least one kilometre (see Plate 3A). It clearly postdates the Coerney Formation and is almost certainly related to the basalt.

FIG. 6.

GEOLOGICAL MAP OF BEYERS VLEY OUTSPAN







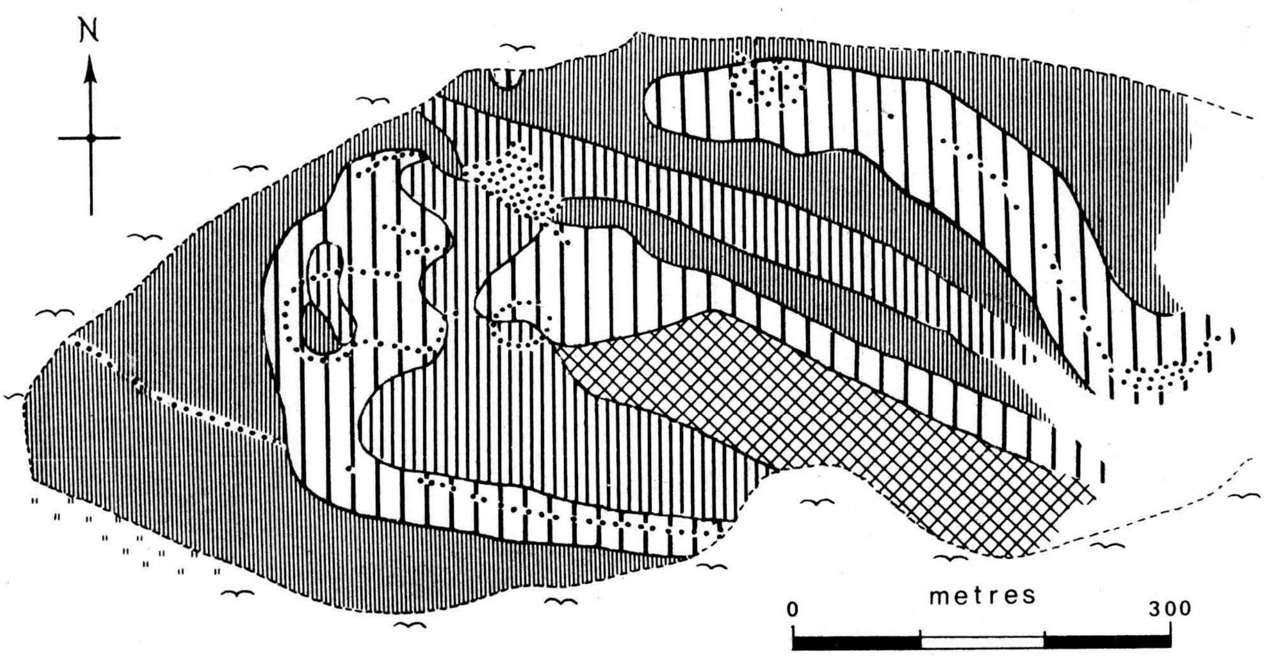
SECTION A–B

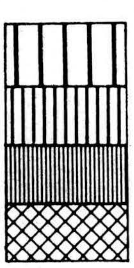
(horizontal scale = vertical scale)

F I G U R E 7

DISTRIBUTION OF BASALT TYPES ON BEYERS VLEY OUTSPAN

(SEE FIGURE 6)





BASALT:

vesicular

less vesicular

rather massive

variable vesicular/massive

······· Dykes & sheets of reddish tuff

c) Tuff

A conspicuous pale red tuff, usually between 0,5 and 3 metres thick, is interbedded with the basalt at many localities. It is particularly common towards the base of the Mimosa Formation.

This tuff has the outward appearance of a very fine-grained, indurated sand=stone. Internal bedding is usually poorly developed or absent. In some cases the thin sheets of red tuff can be followed for several hundred metres laterally (see Plate 1D). On Gorie Laaghte (B) at least four tuff bands occur at different stratigraphic levels within the basalt. Owing to their mode of origin and distribution thin tuff interbeds are usually regarded as important stratigraphic markers. The discontinuous nature of the red tuffs in the Mimosa Formation might indicate short erosional breaks, although no paleosoils or other definite evidence of this kind have been found.

On Enon Mission Station (E) possible vertebrate remains were found in red tuff well above the base of the Mimosa Formation.

The distribution of the red tuff and basalt on Beyers Vlei Outspan (B) seems rather complex (see Figure 6). Three almost circular or horseshoe-shaped hillocks which rise only a few metres above their surroundings and are labelled "volcanic vents" on Map 1, have their outer rims formed by pale red tuffs which dis=play an intrusive relationship with the main body of basalt. Also, as shown by Figure 7, there is a tendency for the red tuff to be associated with more vesicular types of basalt. These little hillocks, which measure several tens of metres across, are clearly former centres of some or other type of volcanic activity. The rocks composing them, especially those near the outer periphery, commonly show near vertical slickensides, possibly as a result of movement caused by eruptive stresses after consolidation of mobile rock material in the upper part of a conduit. Tuff dykes which are sometimes several metres long and up to one metre wide radiate from these vents or plugs. (Plates 1G and 2C). The petrography and possible origin of the tuff dykes on Beyers Vley Outspan will be considered further in Chapters III and V respectively.

Further manifestation of an ancient volcanic centre is evident on Gorie Laaghte (B). This particular spot, which is also indicated as a "volcanic vent" on Map 1, is shown on a larger scale in Figure 8 below. Basalt of the Mimosa Formation and dislocated bodies of red tuff cut across the regional east-west strike of the Suurberg rocks so that it appears as if they pierce the older Coerney tuffs. The red tuff which "floats" in the basalt is intensely shattered; one could speak of a cataclastic tuff-breccia. This shattering is obviously contemporaneous with and a direct result of the development of the vent or diatreme because it is confined to the dislocated tuff bodies.

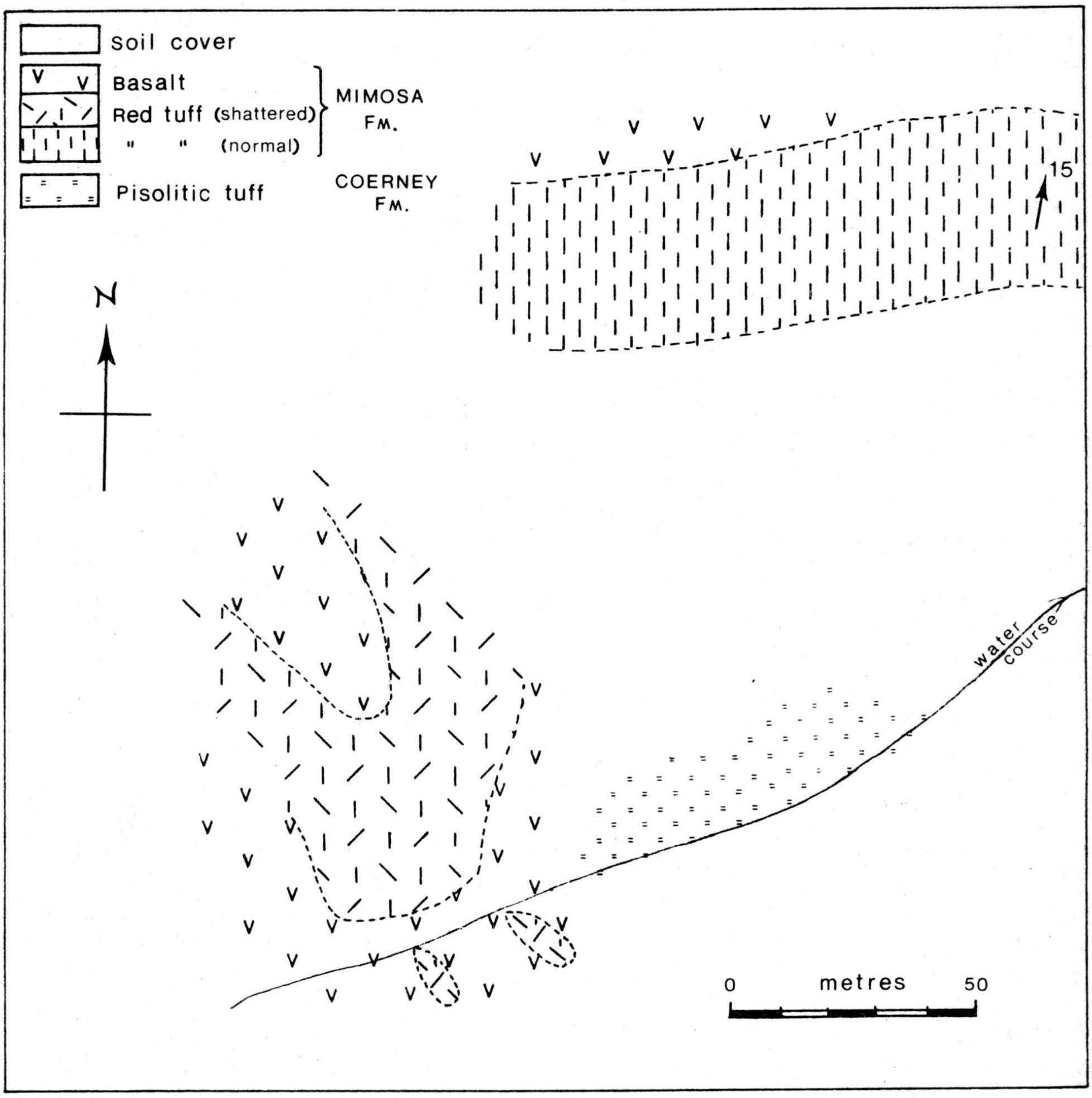


FIG. 8 - Sketch map of outcrops at a "volcanic vent" on Gorie Laaghte.

On Slag Boom (E) and Lot 18 (F) large lenses of greenish, pale orange and pink tuffs, 5 to 15 metres thick and several hundred metres long, are interbedded with basalt. Some of these rocks show faint cross-bedding and are probably either waterlaid or reworked tuffs.

On Rhinoster Hoek no. 2 (B) the basalt is overlain by a reddish tuff with irregular pale grey spots and streaks. A special feature of this tuff is the presence of numerous near vertical tubes which are mostly void and range in diameter from a few millimetres up to 2 cm (see Plate 1C). Individual tubes are usually several tens of centimetres long and tend to branch off or join neigh=bouring tubes. Their walls are generally discoloured.

The origin of the tubes is ascribed to rising gas in the ash immediately after its deposition. This would suggest that the ash was still hot at the time. Guest and Jones (1970) describe identical structures from non-welded rhyolitic ash deposits in central Chili and regard them, together with other evidence, as proof for a pyroclastic flow origin. This aspect will be referred to again in Chapter V.

C. THE UITENHAGE GROUP

Only the Enon Formation and the "Wood Beds" of the Uitenhage Group are present in the mapped area. A survey of the literature reveals that the boundary between these two units is poorly defined and it is generally accepted that they "..... are not strictly successive deposits but were partly contemporaneously formed under different circumstances" (Rogers, 1905, p. 190).

For the purpose of the present investigation a distinction was made purely on lithological grounds. As long as conglomerates were 'prominent' in the succession it was regarded as Enon. This, of course, is a rather subjective criterion and therefore not completely satisfactory.

Problems were for instance encountered in the Bushmans River area and the southern Panhandle. In the case of the former, prominent lenses and streaks of conglomerate are present in a succession which is regarded as "Wood Beds" on some maps. In the south-western Panhandle, on the other hand, some 100 to 200 metres of sandstone and mudstone underlie the first prominent conglomerates.

The present investigation was not directed at an accurate delineation of the Enon-"Wood Beds" transition. Consequently the boundary as shown on Map 1 is only an approximation.

1. The Enon Formation

The Enon is the lowermost formation of the Uitenhage Group. In the mapped area a further distinction can be made between the Basal Enon Sandstone and the Enon Conglomerate proper.

a) The Basal Enon Sandstone

It has already been indicated above that fairly extensive beds of finer clastics make up the basal portion of the Enon locally. These beds form a lithological unit which, in this paper, is informally being referred to as the Basal Enon Sandstone. This unit is restricted to the southern and western parts of the Kirkwood Panhandle where it overlies mainly rocks of the Suurberg Group (see Map 1). Seen as a whole, the sediments of the Basal Enon Sandstone tend to become coarser grained when followed from east to west along their present outcrop area. Upwards they grade into Enon conglomerates.

The Basal Enon Sandstone unit consists mainly of thinly bedded sandstone, smaller amounts of reddish and greenish mudstone, and subordinate small-pebble conglomerates, including tuff conglomerate. The sandstones are commonly trough cross-bedded while others have well developed primary current lineations on bedding planes. The flow directions appear to be extremely variable within short distances.

The thickness of the Basal Enon Sandstone at various localities is shown in

Figure 2, while Figure 5 provides a good example of lithological variations within this unit.

b) The Enon Conglomerate

The Enon Conglomerate is usually several hundred metres thick and enjoys maximum development in the Kirkwood-Paterson area. It comprises principally conglomerate, with subordinate sandstone and mudstone. The nature of the conglo= merate varies slightly from place to place but it is typically closely packed, poorly sorted, and largely unbedded. The clasts of fairly well rounded ortho= quartzite, sandstone, shale, and vein quartz are predominantly pebble- and cobble-sized although boulders up to 50 cm diameter, and slightly more, are found sometimes. The matrix is usually sandy to argillaceous. Red iron oxides and secondary silica act as cementing agents.

The general absence of tillite pebbles in the Enon is remarkable seeing that Dwyka tillite crops out immediately to the north of the area; in some localities, as for instance Enon Mission Station (E), it even underlies the Enon conglomerate. This absence of tillite inclusions was already noted by Rogers (1906, p. 17). It is particularly striking in some poorts, e.g. the Sundays River Poort west of Kirkwood, where tillite pebbles occur together with quartzite pebbles in the modern river bed which cuts through the Enon conglomerate.

Although very rare, a few pebbles of amygdaloidal basalt, agate, and tuff were found in the Enon of the Kirkwood-Paterson and Bushmans River areas. Judged by megascopic appearance the pebbles had almost certainly been derived from the Suurberg Group. It is notable, however, that these inclusions were found strati= graphically some hundred metres and more above the base of the Enon.

In the Bushmans River area the Enon conglomerate differs in some ways from that of other places. Fossil wood and agate pebbles are relatively common and even tillite pebbles are occasionally found.

A very useful characteristic of the Enon conglomerate is the almost parallel vertical fractures which cut through matrix and clasts alike (Plate 2D). These fractures are always best developed near the base of the Enon. The same phenomenon is present in the Slagboom Formation and serves to distinguish these Mesozoic rocks from Cainozoic conglomerates and gravels.

Imbrication of clasts is well developed in the Enon conglomerate which crops out along the northern margin of the Algoa Basin. However, it was found, practi= cally without exception, that the AB-planes (planes of maximum projection) of the clasts dipped downstream, as deduced from cross-bedding, instead of upstream which is normally the case with current transported objects.

Krumbein (1940) noted pebble imbrication in flood gravels that changed from a downstream direction near the head of a canyon to normal upstream imbrication farther downstream. He suggested that the anomalous downstream imbrication had been the result of local causes like reverse currents set up on the lee side of

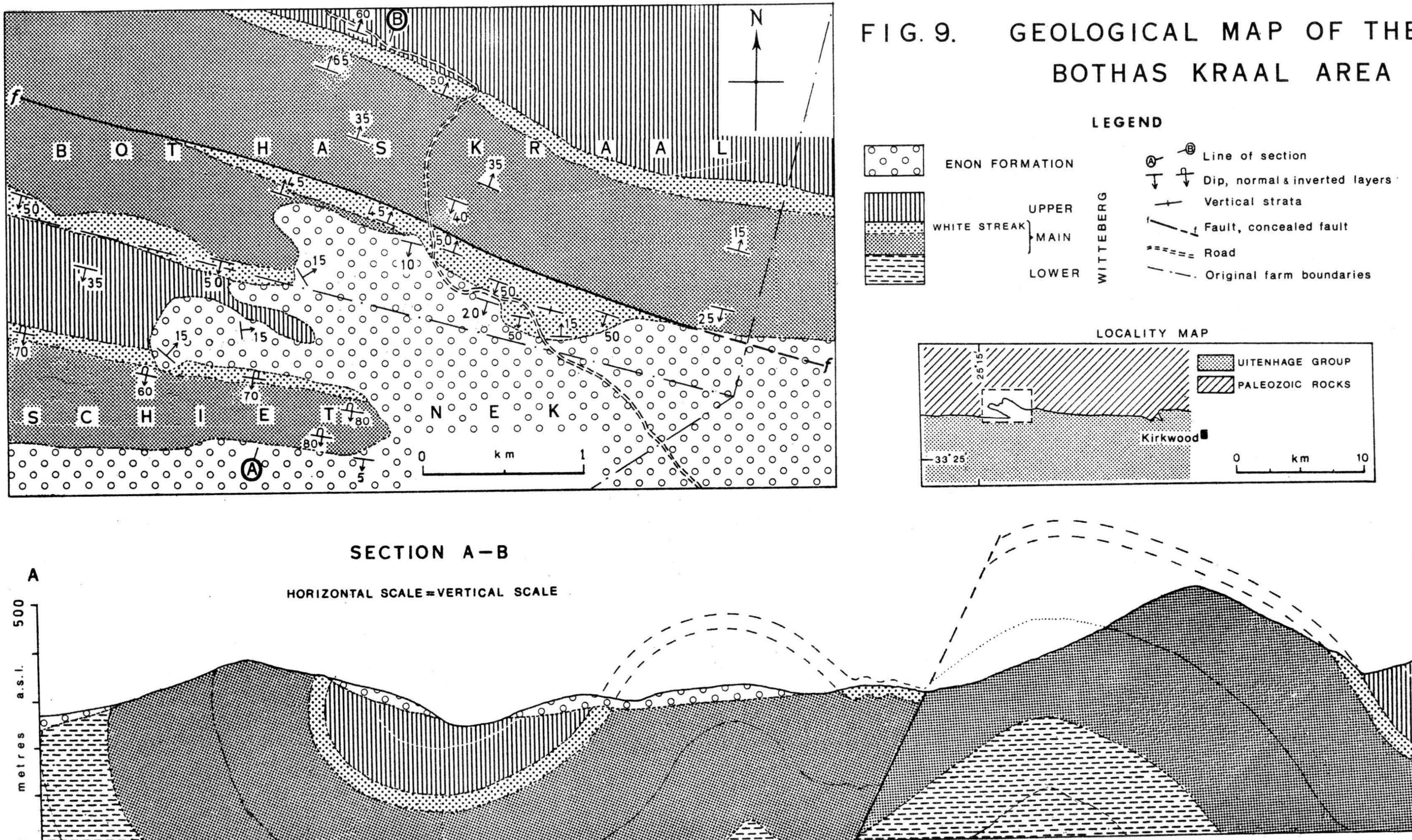


FIG. 9. GEOLOGICAL MAP OF THE BOTHAS KRAAL AREA

large boulders (Krumbein, 1942, p. 1389). Further, Potter and Pettijohn (1963, p. 35) point out that if inclination of the AB-plane is measured from the horizontal rather than the foreset plane in the case of cross-bedded units apparent downcurrent imbrication might be found.

Because of the prevalence of this phenomenon in the Enon conglomerates it is most likely that neither of the above-mentioned factors is applicable. The author has not investigated this problem adequately in the field to be able to advance a well-founded explanation. It is tentatively suggested that the initial slope of the depositional surface played a role. A relatively steep inclination would probably allow clasts in the upper parts of a moving mass of debris to slide over underlying clasts so that the former would come to rest in a position of downstream imbrication.

On Botha's Kraal (D) a bedded quartzite breccia which grades vertically into conglomerate occurs at the base of the Enon. This breccia is obviously derived from a nearby pre-Enon fault (see Fig. 9). It reminds of the Slagboom breccia but for its bedded nature and the complete absence of tuff material.

c) General

The lower contact of the Enon Formation with the Suurberg Group is sedimentary in practically all the exposures seen. On Erenkrons Poort (A), and also in the Bushmans River area, the Enon overlies tuffs of the Coerney Formation. At the first mentioned locality the contact plane is erosional and somewhat irregular but in the Bushmans River area, where the Enon appears to be transgressive, no clear contacts with the tuffs are exposed.

The lower contact with the Mimosa basalt is remarkably straight and apparently conformable as seen at several places (Plates 2F, 3B & 3E). However, it is usually hard to find a good exposure of the actual contact plane. At two localities, Erenkrons Berg (A) and the farm Coerney (F), where it was laid bare, the basalt immediately below the Enon was too weathered to allow definite deductions. It is uncertain whether this weathering is post-depositional or not.

Between Erenkrons Berg (A) and Enon Mission Station (E) the Enon overlies Paleozoic rocks with a pronounced angular unconformity. This unconformity is well exposed at the following localities: Haas Poort (B), along the western bank of the Haaspoortspruit; approximately one to two kilometres W.N.W. of Sapkamma railway siding; on Schiet Nek (D); Courans Drift (D), along the western bank of the Sundays River; about three kilometres due west of Kirkwood; and also on Aluin Krantz West (G) (see Plates 3C & 3D).

Attention has already been drawn above to the paucity of basalt and tuff clasts in the lower parts of the Enon Formation. An exception is found along the western extremity of the Kirkwood Panhandle where inclusions of tuff are common in the basal rocks of the Enon Formation and in places proper tuff conglomerates have developed. But even here basalt inclusions are invariably absent although the Enon is often directly underlain by this rock type (see Fig. 4).

This peculiarity is all the more remarkable in the light of the situation which exists under apparently equivalent circumstances elsewhere. Frankel (1960) describes a Lower Cretaceous conglomerate, with abundant basalt clasts, which over=laps on to steep pre-Cretaceous slopes in Natal. These slopes are formed by in=clined early Jurassic basalt sheets which clearly acted as source for the clasts in the conglomerate.

It would appear then that the absence of basalt clasts in the lowermost Enon of the Algoa Basin could be ascribed to the non-availability of basalt at the time when deposition of the Enon started, rather than subsequent destruction of such clasts. The possible reasons for this will be pursued further in Chapter V.

2. The "Wood Beds"

No particular attention was paid to this unit. In the southeastern part of the Kirkwood Panhandle, where this unit occurs along the inner margin of the Algoa Basin, it comprises mainly poorly exposed reddish, greyish and bluish mudstones. In the Bushmans River area sandstones predominate, with mudstone less abundant. Subordinate lenses of conglomerate and grit are present, as well as abundant fossil wood.

D. THE CAINOZOIC DEPOSITS

The younger units, comprising the Alexandria Beds, terrace gravels at various levels, and more recent deposits have no direct bearing on the earlier depositional history of the Algoa Basin. They are only mentioned briefly because they commonly cover the older rocks and are shown as such on Map 1.

A tentative boundary between Tertiary and Quaternary gravels was drawn based on inferred ancient sea levels (Du Toit, 1954, p. 464). All terrace gravels lying higher than 75 metres above the nearest prominent valley floor or drainage line were regarded as "Tertiary" and those below as "Quaternary". The Tertiary gravels were further subdivided into "High Level" and "Intermediate Level" gravels depending on whether they occurred higher or lower than 150 metres above the valley floor.

The shortcomings of the above subdivision are fully appreciated because it does not take into consideration paleoslopes and later downwarping. On the other hand, until more reliable data become available, this distinction provides a working basis. Also, it should be mentioned that the three episodes of Cainozoic tilting and warping determined by Ruddock (1968) in the southern Algoa Basin have east-west axes which lie south of 33°30' latitude, i.e. south of the mapped area.

The Tertiary terrace gravels, as defined above, consist essentially of orthoquartzite pebbles, with Dwyka tillite pebbles conspicuously absent. In the southern Panhandle area they have almost certainly been derived from the south.

A few pebbles with Bokkeveld fossils have been noted amongst others.

The younger river terrace gravels and alluvium usually bear a clear record of the lithology of their source areas. Tillite pebbles are, for instance, quite common in modern stream beds which drain areas underlain by Dwyka rocks.

III LABORATORY INVESTIGATIONS

A. OBJECTS AND METHODS

Only rock samples of the Suurberg Group and the lowermost portion of the Enon Formation were subjected to further laboratory investigation. Fresh, representative samples were selected as far as possible. For the Cape and Karoo rocks it was assumed that petrographic results from regional studies (amongst others: Theron, 1970; Loock, 1967; De Villiers and Wardaugh, 1962) would hold good for the area under discussion.

The primary objective of the laboratory work was to determine the relative amounts of important and characteristic mineral components present in the Suur= berg and basal Enon rocks. This would not only permit correct rock classifi= cation, but should also throw more light on stratigraphic relationships and the possible origin of the rocks. On the other hand, due to the many rock types and minerals involved, a too elaborate mineralogical investigation was avoided, lest the present research extend beyond its limits.

With these objectives in view, methods were employed which would allow rapid identification of a larger number of grains, rather than more sophisticated techniques that would be too time consuming in relation to the higher accuracy attained. Optical axial angles were measured by the method of Tobi (1956) and refractive indices by the immersion method in yellow light. Plagioclase compo= sitions were determined by extinction angles (cleavage fragments; albite twins; combined Carlsbad-albite twins) and refractive indices using the various identi= fication tables in Deer et al. (1966).

Volumetric composition of rocks was determined with a Leitz integrating stage. Two or more differently orientated thin sections per sample were inte= grated along equally speced traverses over a total distance of at least 15 centi= metres.

Thin sections of clastic rocks were stained with sodium cobaltinitrite (Deer et al., 1966, p. 311) to simplify distinction between K-feldspar and quartz or untwinned plagioclase. Unfortunately thin sections of tuffs gene= rally produced disappointing results. The matrix usually took on a bright yellow colour (and should therefore contain some potassium) but phenoclasts[+] were not significantly affected although many of them were known to be K-feldspars. As a check the same staining method was tried on granite, with good results. This experience seems to confirm the conclusion that alkali feldspars, including sanidine and anorthoclase of many volcanic rocks, do not respond as readily to the potassium-stain as do alkali feldspars of plutonic rocks (Chayes and Zies, 1961, p. 172-173).

The problem of distinguishing between quartz and untwinned feldspar in tuff thin sections was partly overcome by using high power magnification so that interference figures could be quickly obtained when necessary. At the same time cleavage traces and degree of alteration could be noted more carefully.

B. PETROGRAPHY

The Suurberg Group and the overlying Enon consist of four main rock categories — rudite, tuff, basalt, and arenite (excluding tuff) — which are treated separately.

1. Rudites

Because of their megascopic character the petrography of the conglomerates and breccias of the Enon and Slagboom Formations has already been treated in the section on field geology.

The clayey matrix of the Slagboom breccia from Drie Kuilen (A) was thermally investigated. The D.T.A. curves showed good repeatability and a pronounced endothermic peak developed around 180°C. Of the commonly known clay mineral groups smectite appears to be the most likely to have produced this peak (Grim, 1968, p. 314). Smaller peaks which developed at higher temperatures (e.g. around 570°C and 850°C) were probably due to quartz and other "impurities" and did not confirm the presence of any specific clay mineral.

Agglomerate is occasionally found in the Coerney Formation. Thin section study reveals that the tuff fragments which make up the rock differ in no way from other pale orange and pink tuffs of the same formation. The agglomerate is cemented largely by zeolite.

2. Tuffs

The Coerney Formation consists essentially of tuff while subordinate tuff layers also occur interbedded with the Mimosa basalt. The most important

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[+]Most other authors use the term "phenocryst" for the crystal fraction which lies embedded in the fine groundmass of tuffs. The term "phenoclast" is preferred here because pyroclastic rocks, although largely endogenetic, are clastic and not crystalline.

petrographical characteristics of the various types of tuff are presented in Table 3. These will be briefly discussed under the headings "Texture" and "Composition".

a) Texture

A distinctive feature of all the tuffs in thin section is their "porphyritic" nature. Relatively larger broken crystals, glass shards and, to a lesser extent, lithic fragments are embedded in a matrix of much finer material which makes up between 50% and 90% of the whole rock (see Plate 5 A & B). The resemblance in thin section with greywackes is often striking. However, the presence of glass shards and, occasionally, some small pumice fragments leaves little doubt as to the pyroclastic origin of the rocks dealt with here.

Because the tuffs are rather well consolidated, mechanical size analyses were not practicable. Some idea of the sizes of the phenoclasts was obtained by measuring 100 randomly selected grains per slide. The phenoclasts generally fall in the silt to medium sand size grades of the Wentworth scale. Occasionally larger, up to pebble-size , accidental lithic fragments, mostly quartzite, are found scattered sparsely through some tuffs, especially those near the base of the Coerney Formation.

Phenoclasts are usually fragmented+ while others are rounded and sometimes slightly embayed. In cross section the glass shards are mostly platy, crescentic or Y-shaped representing broken walls of either individual or clusters of gas bubbles. In some tuffs the outlines of shards are enhanced by a thin rim of red iron oxide. Where many glass shards are present the tuffs have a typical vitro= clastic texture (Plate 4 C).

In some of the tuffs glass shards are feebly to distinctly welded so that they are flattened and distorted. In extreme cases of welding, shards have been drawn out to produce a pseudo-flow structure which is visible under the micro= scope (Plate 5 E). The glass of the welded tuffs are partially or completely devitrified and often result in the characteristic axiolitic texture of ash-flow tuffs (Ross & Smith, 1961, p. 37) (see also Plate 5 F). The fact that certain Suurberg tuffs are welded was only discovered under the microscope.

b) Composition

The tuffs consist mainly of a groundmass of presumably volcanic dust and glass which is partly or wholly devitrified and optically irresolvable. Occasionally the matrix is also partly zeolitized or "impregnated" by green celadonite (Plate 5 E).

+"Fragmented" crystals have been separated into parts whereas "cracked" ones still retain their crystal form.

COMPOSITION OF SUURBERG TUFFS (% by volume) — PHYSICAL PROPERTIES OF TUFFS

Sample	Quartz	K-feldspar	Plagioclase	Biotite	Hornblende	Heavy minerals	Lithics	Glass shards	Matrix	Colour	Refr. index (± 0,005) Matrix	Refr. index Glass shards	S.G. (± 0,01)	Size of crystals (mm) Max.	Size of crystals approx. mode	Welding	Rock name	Unit
1	19,7	13,5	9,2	tr	-	0,9	3,2	p	53,5	5R5/4	1,500	-	2,21	0,25	0,08	-	tubular tuff	MIMOSA FORMATION
2	11,0	3,7	8,2	0,4	-	0,1	0,3	14,1	62,2	5Y7/2	-	1,495	2,29	0,26	0,08	-	waterlaid tuff	
3	12,3	9,1	8,9	-	-	0,4	0,2	3,5	65,6+	5GY7/1	1,510	1,515	2,47	0,22	0,07	distinct	green tuff	
4	10,8	4,2	7,0	0,4	0,2x	1,4	0,1	p	75,9	5YR5/2	1,525	n.d.	2,39	0,17	0,07	slight	dyke pale red tuff	
5	7,3	7,7	7,9	1,4	0,3x	1,2	0,2	p	74,0	5YR6/2	1,510	n.d.	2,30	0,15	0,06	slight		
6	16,4	6,9	4,6	0,6	0,2	1,1	-	p	70,2	10R6/2	1,515	n.d.	2,26	0,17	0,06	distinct	sheet pale red tuff	
7	8,2	11,6	7,6	tr	-	1,4	0,9	p	70,3	10R5/4	1,505	n.d.	2,34	0,16	0,07	slight		
8	10,7	5,8	10,2	tr	-	1,4	0,5	p	71,4	10R5/4	1,510	n.d.	2,34	0,12	0,06	slight		
9	16,8	7,0	5,0	tr	-	0,8	-	p	70,4	5YR6/2	1,525	n.d.	2,28	0,24	0,08	distinct		
10	15,9	1,4	7,3	0,2	-	1,2	0,4	p	73,6	10R7/4	1,510	n.d.	2,26	0,13	0,07	slight		
Mean	12,3	6,3	7,1	0,4	0,1	1,2	0,3	p	72,3		1,515	-	2,31	0,16	0,07			
11	0,3	3,9	1,5	1,5	-	0,3	-	12,5	80,0	5Y8/1	1,505	1,500	2,17	0,16	0,05	-	pale whitish tuff	COERNEY FORMATION
12	1,0	3,0	2,6	0,9	-	0,3	-	6,7	85,5	5Y8/1	1,505	1,505	1,98	0,09	0,04	-		
13	0,3	4,6	1,9	0,4	-	0,2	0,5	p	92,1	5Y8/1	1,495	n.d.	1,88	0,15	0,05	-		
14	1,6	6,2	2,2	2,6	-	0,6	-	(p)	86,8	5Y8/1	1,525	n.d.	2,13	0,28	0,06	-		
15	0,1	1,7	1,1	0,2	-	0,1	-	23,4	73,4	N8	1,485	1,485	2,10	0,25	0,08	-		
Mean	0,7	3,9	1,8	1,1	-	0,3	0,1	10,7	81,4		1,505	1,497	2,05	0,19	0,06			
16	9,7	3,3	4,6	1,1	tr	0,4	-	(p)	80,9	10YR6/6	1,490	n.d.	1,93	0,18	0,09	-	pisolitic tuff	
17	9,2	3,0	4,4	0,3	-	0,2	4,3	3,2	75,4	10YR7/4	1,510	1,505	2,26	0,34	0,06	-		
18	9,4	3,3	5,0	0,2	-	0,3	1,2	1,8	78,8	5YR6/4	1,520	n.d.	2,24	0,22	0,09	-		
Mean	9,4	3,2	4,7	0,5	-	0,3	1,8	2,5	77,6		1,507	-	2,14	0,25	0,08			
19	9,6	11,7	10,9	0,4	-	0,5	0,2	p	67,7+	10YR6/2	1,520	n.d.	2,32	0,12	0,05	distinct	pale orange or pink tuff	
20	21,2	9,3	6,2	0,1	tr	0,8	0,6	p	61,7	5YR6/2	1,505	n.d.	2,38	0,15	0,06	distinct		
21	11,5	5,1	7,0	0,5	-	0,9	0,4	12,6	62,0	5YR7/2	1,495	1,495	2,20	0,18	0,09	-		
22	4,8	3,9	3,5	0,9	-	0,2	0,3	p	86,4	10YR8/6	1,510	n.d.	1,95	0,12	0,05	-		
23	7,0	2,9	2,3	0,4	-	0,2	0,2	p	87,0	5GY8/1	1,540	n.d.	2,19	0,24	0,05	-		
24	10,9	7,3	5,1	0,6	-	0,9	4,1	18,3	52,8	5Y8/2	1,510	1,505	2,27	0,65	0,08	-		
25	8,0	7,2	7,9	0,6	-	0,3	1,0	27,6	47,4	5Y8/2	1,500	1,500	2,05	0,17	0,05	-		
26	9,1	1,4	4,8	1,8	tr	1,3	-	8,3	73,3	10YR7/4	1,515	n.d.	2,08	0,15	0,08	-		
27	5,8	7,5	4,8	0,8	tr	0,6	0,4	p	80,1	5YR7/2	1,485	n.d.	2,18	0,10	0,05	-		
28	6,2	2,0	4,0	0,2	tr	0,3	1,0	p	86,3	10R6/6	1,490	n.d.	2,20	0,28	0,07	-		
29	9,9	3,7	7,2	0,3	tr	0,2	1,6	p	77,1	10YR7/4	1,490	1,490	2,13	0,15	0,08	-		
30	5,0	1,3	6,1	0,4	tr	0,3	0,1	p	86,8	5YR6/4	1,515	n.d.	2,19	0,09	0,05	-		
31	5,2	3,8	5,1	1,0	-	0,4	1,2	p	83,3	5YR7/2	1,505	n.d.	2,17	0,18	0,06	-		
32	10,3	4,6	4,5	1,2	-	0,4	tr	4,6	74,4	5YR8/1	1,510	n.d.	2,07	0,19	0,05	-		
Mean	8,9	5,1	5,7	0,6	tr	0,5	0,8	14,3	62,1		1,505	1,498	2,14	0,20	0,06			

x incl. basaltic variety; + incl. celadonite; tr = trace; p = present but not distinguished from matrix; n.d. = not determined

Sample localities:

1 - Rhinoster Hoek no. 2 (B)
2 - Slag Boom (E)
3 - Enon Mission Station (E)
4&5 - Beyers Vley Outspan (B)
6 - Enon Mission Station (E)
7 - De Vlei (A)
8 - Beyers Vley Outspan (B)
9 - Gorie Laaghte (B)
10 - Drie Kuilen (A)
11 - Tover Klip (C)
12 - Erenkrons Poort (A)
13 - Rhinoster Hoek no. 2 (B)
14 - Woodbury (G)
15 - Springbok Vlackte no. 6 (A)
16 - Erenkrons Poort (A)
17 - Unamore (F)
18&19 - Slag Boom (E)
20 - Enon Mission Stn. (E)
21 - Beyers Vley Outspan (B)
22 - Erenkrons Poort (A)
23 - Zand Vlakte (F)
24 - Enon Mission Stn. (E)
25 - Brandt Koppen (C)
26 - Matjesgoed Fontein (B)
27,28,29&30 - Springbok Vlackte no. 6 (A)
31 - Gorie Laaghte (B)
32 - The Retreat no. 1 (G)

The phenoclasts are mainly of quartz, sanidine, sodic plagioclase, biotite, hornblende, and various heavy minerals. In addition, a very small percentage of lithic fragments is present, as well as variable amounts of glass shards.

Quartz occurs as long slivers and broken crystals with irregular outlines, some of which are slightly resorbed and embayed. The mineral is always fresh and commonly contains small specks or inclusions. Some quartz crystals are probably strained, as indicated by biaxial interference figures with 2V's up to 5°. Accidental quartz crystals, if present, could not be distinguished.

Potash feldspars are almost exclusively sanidine. A few specimens with axial angles in the order of 40° - 50° are probably anorthoclase. Microcline occurs in a few samples but forms only a very small fraction of one per cent of the total number of feldspars present.

Sanidine crystals are characteristically tabular, when not fragmented, with a smooth glassy surface and occasionally a very pale brownish colour. The optic plane lies perpendicular to (010) with $2V_{\alpha}$ varying between 5° and 25°. This variation in axial angle for different sanidine crystals from the same rock specimen appears to be no unusual feature (MacKenzie and Smith, 1956, p. 407). Relatively few crystals are twinned, usually according to the Carls= bad law, and no definite zoning is visible under the microscope.

Plagioclase grains are either polysynthetically twinned or untwinned. The latter are often difficult to distinguish with certainty from sanidine or quartz, but usually the degree of alteration provides a useful clue when interference figures cannot be obtained. Also, many plagioclase crystals have normal conti= nuous zoning. The plagioclase is mostly oligoclase with average composition between An_{20} and An_{30}. The plagioclases will be discussed further in a later section.

Biotite and hornblende occur as shreds in most of the tuffs. Basaltic hornblende, distinguished from ordinary hornblende by its higher refractive indices and characteristics pleochroism (γ = dark reddish brown, α = pale yellow), is present in the tuff dykes from the volcanic vents on Beyers Vley Outspan (B). No pyroxene crystals were noted, although Haughton (1935, p. 20) mentions the presence of augite in tuff close to a basalt contact. These were probably accessory fragments derived from the basalt.

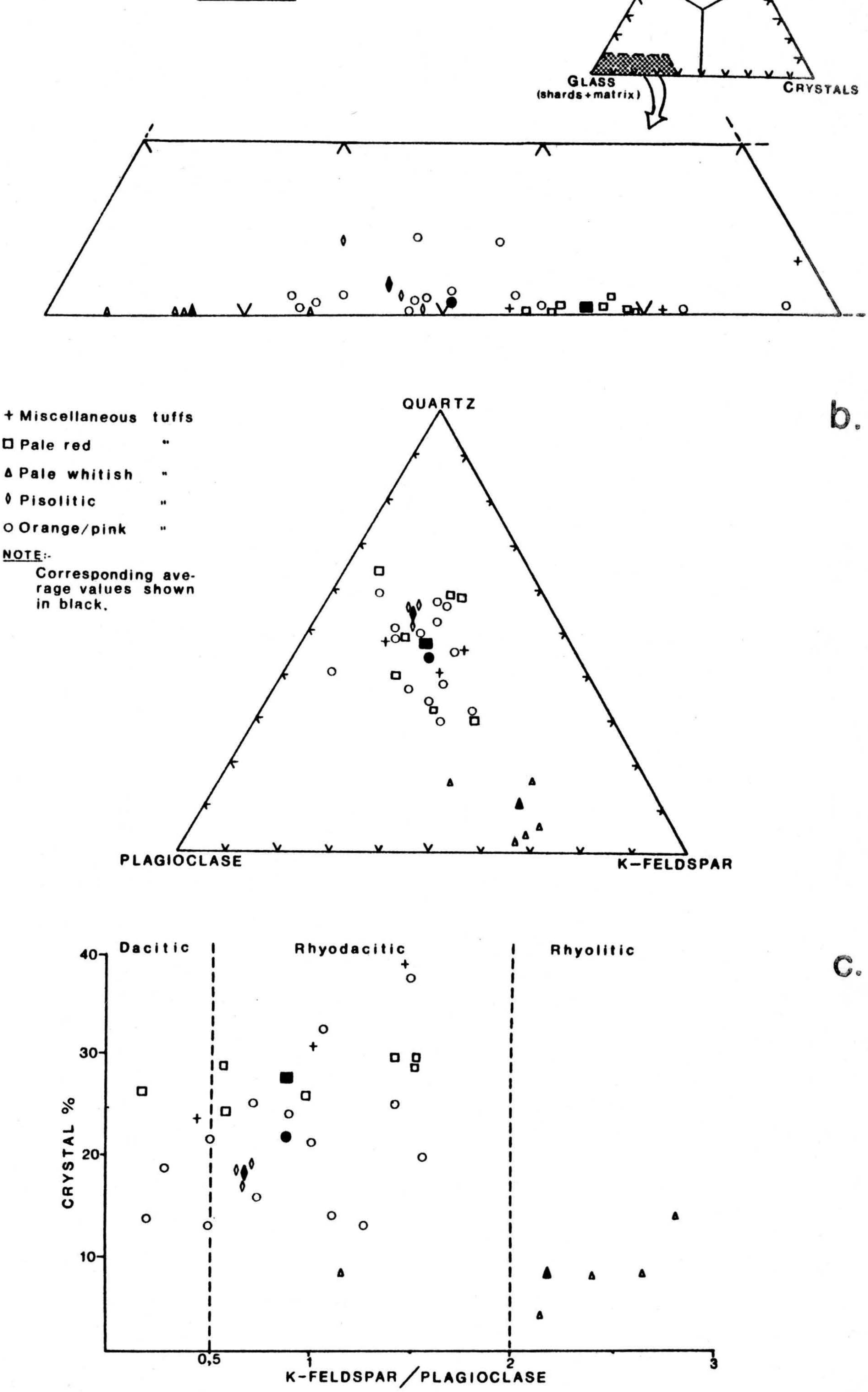
The commonest heavy minerals in the tuffs are magnetite and ilmenite. Zircon, apatite, epidote, garnet and sphene are scarcer.

The few rock fragments in the tuffs are mostly derived from the Cape Super= group and include siltstone, quartzite, and shale. Accessory tuff and basalt fragments are much rarer and usually occur in the Mimosa tuffs.

Glass shards are present in practically all tuff thin sections. Yet, they are not always easily distinguished from the groundmass when both are devitrified or when welding is pronounced. The degree of devitrification varies from incipient to complete, but in some samples glass shards are unaltered. These last mentioned shards have refractive indices ranging between 1,485 and 1,515 which would correspond to a SiO_2 composition of between 76% and 66% respectively (George, 1924).

FIGURE 10

COMPOSITIONAL VARIATION OF (a) THE SUURBERG TUFFS, AND (b & c) THEIR CRYSTAL FRACTION



A few small pumice fragments are also included with the glass shards in Table 3.

Celadonite is abundant in the matrix of certain green tuffs (8,2% of sample 3 in Table 3 is composed of this mineral). In some of the other tuffs traces of celadonite are sparingly disseminated through the matrix. A few small grains allowed its optical properties to be determined:

$2V_{\alpha} \geq 0^{o}$	Pleochr. : γ = deep bluish green,
$\gamma = 1,615$ $(\pm 0,003)$,	α = yellowish green;
$\alpha = 1,585$ $(\pm 0,003)$;	Dispersion : strong $r > v$

Certain of the compositional properties listed in Table 3 are also presented diagrammatically in Figure 10. The following features are apparent:

i) All the Suurberg tuffs fall in the category "vitric tuff".

ii) Except for the pale whitish tuffs, which appear to form a distinct group, no systematic trends emerge for the various tuffs.

iii) When only their feldspar crystal content is considered the pale whitish tuffs seem to be relatively more acidic. It is interesting, however, that quartz crystals form a comparatively lower proportion of the phenoclasts of this tuff type.

iv) From the crystal fraction of the tuffs it is deduced that they vary in composition from dacitic to rhyolitic with the majority corresponding to rhyodacites.

Peterson and Roberts (1963) suggested a possible inverse relationship between the crystal content and silica content of certain tuffs. This tendency is not very pronounced in the Suurberg tuffs although the pale whitish tuffs, which in= variably have a low crystal content, appear to be the most acidic.

The experience of other workers has shown that the groundmass of tuffs is richer in SiO_2 than the crystal fraction, which is to be expected in the light of the normal crystallization sequence in magmas. Consequently the tuffs as a whole should normally be more acidic than indicated by mere inspection of the crystal fraction.

3. Basaltic Rocks

a) Basalt

Under the microscope many of the basalts exhibit a microporphyritic texture. The microphenocrysts, consisting of plagioclase laths and less frequently stubby clinopyroxene grains, are embedded in a dark brown, almost opaque base of altered glass and crystallites.

The composition of basalt samples is shown in Table 4. Sample numbers indicate, more or less, progressively higher positions above the base of the Mimosa Formation. The minerals which fill vesicles were excluded from the count. Although Table 4 reflects large variations in the degree of crystallization of the

TABLE 4

COMPOSITION OF SUURBERG BASALT (% by volume)

	Plagio= clase	Clino= pyroxene	Olivine (*)	Ore	Ground= mass	Remarks
15	42,5	29,4	2,8	5,5	19,8	rather massive
14	49,4	24,3	0,1	5,6	20,6	rather massive
13	48,8	27,7	2,8	5,6	15,1	amygdaloidal
12	50,1	34,5	1,2	4,5	9,7	amygdaloidal
11	44,7	28,8	2,5	3,9	20,1	massive
10	11,1	5,9	-	m	83,0	amygdaloidal
9	37,3	4,1	-	m	58,6	slightly amyg.
8	45,1	25,3	tr	4,6	25,0	amygdaloidal
7	13,5	2,4	0,9	m	83,2	amygdaloidal
6	45,6	29,3	-	6,1	19,0	rather massive
5	44,0	27,6	0,1	5,1	23,2	rather massive
4	18,6	5,2	1,3	m	74,9	amygdaloidal
3	42,9	15,7	2,5	6,8	32,1	amygdaloidal
2	36,1	10,4	1,1	m	52,4	amygdaloidal
1	47,9	32,1	4,1	4,5	11,4	slightly amyg.
Mean	38,5	20,2	1,3	5,2	34,8	

tr = trace ; m = included with groundmass

(*) Iddingsite and serpentine pseudomorphous after olivine.

Sample localities:

1 - Gorie Laaghte (B)
2 - Rhinoster Hoek no. 2 (B)
3 - Gorie Laaghte (B)
4 - Thornleigh (F)
5 - Erenkrons Poort (A)
6 - Drie Kuilen (A)
7 - Springbok Vlackte no. 2 (B)
8 - Beyers Vley Outspan (B)
9 - Drie Kuilen (A)
10 - Kremlin (E)
11 - Zand Vlakte (F)
12 - Matjesgoed Fontein (B)
13 - De Vlei (A)
14 - Beyers Vley Outspan (B)
15 - Slag Boom (E)

lava the broad mineralogical composition appears to be consistent both laterally and vertically.

Twinned and zoned plagioclase is the major constituent of the basalt. In some thin sections crystal outlines display considerable resorption with little or no signs of repair. Otherwise, feldspars from fresh samples show little alteration. The composition of plagioclase from all thin sections falls in the range An_{56} - An_{72}, but calcic labradorite predominates. The plagioclase appears to have crystallized before the pyroxene because the latter often encloses euhedral plagioclase grains. Also, it will be noted from Table 4 that samples with a high proportion of chilled groundmass have a particularly low pyroxene : plagio=clase ratio.

Clinopyroxenes occur in subhedral grains which are generally smaller than the feldspar crystals of the same rock. Zoning, twinning, and slight resorption of crystals are not uncommon. Both augite and pigeonite are present, but they are not readily distinguished in thin section. However, by investigating the interference figures of a number of randomly selected grains it was found that augite predominates in all samples. Also, the pigeonite grains tend to be smaller. In fresh samples the pyroxenes are practically unaltered.

Although the optical properties of the augite are somewhat variable the following may be regarded as typical values for augite of the Mimosa basalt:

$$\begin{aligned} \gamma - \alpha &= 0{,}030 \quad (\pm\, 0{,}003) \\ \beta &= 1{,}690 \quad (\pm\, 0{,}003) \\ 2V_\gamma &= 46^\circ \quad (\pm\, 4^\circ) \end{aligned}$$

These values would plot well within the augite field (Deer et al., 1966, p. 125) although somewhat displaced towards the endiopside boundary.

The pigeonite has variable optic axial angles which are usually smaller than 10° but in some grains may be as much as 25°. The refractive indices for this mineral are only slightly higher than the corresponding values for augite.

Although no fresh olivine is present, pseudomorphs after this mineral occur in some slides (Plate 6C). Apparently olivine was irregularly distributed throughout the basalt but never attained major proportions. The principal mineral replacing olivine is iddingsite which is characterized by its slight pleochroism from deep red brown (γ) to deep orange brown (α), high birefringence, $\beta \geq 1{,}72$, $2V_\gamma = 30^\circ - 40^\circ$, and strong dispersion ($r > v$).

Small black grains of ore are sometimes subhedral to euhedral, which assisted in their identification as magnetite and ilmenite. They mostly occur in the matrix as finely disseminated grains, but sometimes they are also en=closed by plagioclase and pyroxene.

Small rounded grains of apatite and, more rarely, zircon are present as very minor constituents. No quartz, alkali feldspars, orthopyroxenes or amphiboles were noted among the primary minerals of the basalt.

Secondary minerals which fill many gas holes and seams in the basalt in= clude celadonite, zeolites, chalcedony, and calcite.

Celadonite has a distinctive bluish green colour and occurs in radiating fibres showing distinct pleochroism. The optical identification (see p. 28) was verified by X-ray diffraction (J.D.T. Otto). Where present, celadonite invariably lines the outer rim of gas holes followed by other secondary minerals towards the centre of the cavity (Plate 6 C).

Heulandite appears to be the commonest zeolite associated with the basalt, but analcime was also noted in a few thin sections. No further systematic study was made of the zeolites and their distribution.

b) Dolerite

In contrast to the basalt, the dolerites of the Suurberg Group are holocrystalline. Subophitic and ophitic textures are typically developed; large augite grains which measure between 0,5 and 2,5 mm enclose smaller plagio= clase laths (Plate 6 D).

TABLE 5

COMPOSITION OF SUURBERG DOLERITES (volume %)

Sample	Plagio= clase	Clino= pyroxene	Ore	Intersertal yellow-brown & green minerals	Locality
1	48,5	37,9	3,0	10,6	Enon Msn. Stn. (E)
2	47,1	41,4	2,5	9,0	Slag Boom (E)
3	48,9	35,9	2,8	12,4	Slag Boom (E)
Mean	48,2	38,4	2,8	10,6	

Table 5 above presents the volumetric composition of three dolerite samples. Unfortunately direct comparison between Tables 4 and 5 is not possible on account of the high proportion of indeterminate groundmass in the former. Still, it would appear that, mineralogically, the basalt and the dolerite are rather similar.

The plagioclase of the dolerite is principally calcic labradorite and op= tically there is practically no difference between the clinopyroxenes of the dolerite and those of the basalt.

A special feature of the dolerite is the occurrence of interstitial yellow-brown and greenish minerals, in the form of platelets and radiating bunches, between quite fresh plagioclase and augite. In only a few instances are there indications of slight replacement of plagioclase and pyroxene by these minerals. Two distinct minerals appear to be present. The green one is optically similar to the celadonite found in the amygdaloidal basalt. The yellow-brown mineral has optical properties:

Pleochroism: α = brownish yellow, γ = dark yellowish brown;
$2V_\alpha \geq 0^0$; $\beta \simeq 1{,}580$; $\gamma - \alpha > 0{,}03$;

but could not be identified conclusively. It appears to belong to the sheet silicates, possibly stilpnomelane. Wise and Eugster (1964, p. 1074) mention a possible interrelationship between celadonite and stilpnomelane. The two minerals probably formed late in the crystallization history of the dolerite.

4. Sandstone

The Basal Enon Sandstone which overlies the Suurberg volcanics in the southern Kirkwood Panhandle is the only Mesozoic sandstone unit with direct bearing on the problems of the present paper.

Texturally these sandstones appear rather variable under the microscope. Factors like sorting, rounding, and grain size all differ considerably from place to place.

More importance was attached to the composition of the sandstones (Table 6).

TABLE 6

COMPOSITION OF BASAL ENON SANDSTONES (volume %)

Sample	Quartz	Alkali feldspar	Plagio= clase	Mica	Heavy Minerals	Lithics	Glass shards	(Clayey) Matrix
1	29,1	4,2	1,3	0,2	0,1	30,6	-	34,5
2	20,2	12,6	6,4	tr	0,2	34,1	0,9	25,6
3	16,9	12,2	7,2	tr	0,2	33,4	1,4	28,7
4	24,2	4,3	2,1	0,1	0,2	50,3	-	18,8
Mean	22,6	8,3	4,2	0,1	0,2	37,1	0,6	26,9

Sample Localities:

1 - Rhinoster Hoek no. 2 (B)
2 - Erenkrons Poort (A)
3 - Erenkrons Poort (A)
4 - Drie Kuilen (A)

Rock fragments predominate and consist chiefly of quartzose siltstone, shale, quartzite, and tuff (in decreasing order) which, with the exception of the tuff fragments, have clearly been derived from the Paleozoic "basement". Second in abundance is the matrix which consists of an indeterminate clayey paste.

Two hundred randomly selected untwinned feldspar grains were investigated conoscopically. Of these 56% proved to be sanidine while the remainder re=

presented other varieties of alkali feldspar and untwinned plagioclase. Of all feldspars counted less that 0,5% was microcline. Twinned plagioclase grains were mostly oligoclase.

Further noteworthy components of the sandstones are glass shards, many of which are undevitrified (Plate 5 C), and mica. The latter is conspicuously rare although it is fairly prominent in the rock fragments. The matrix might contain cryptocrystalline mica or alteration products of this mineral.

C. COMPARATIVE MINERALOGY

Some further work was done on the plagioclase and heavy mineral content of the Suurberg and basal Enon clastic rocks with the object to obtain additional comparative data, as well as evidence regarding the source of these rocks.

1. Plagioclase

One thousand randomly selected twinned plagioclase grains were grouped by means of refractive indices into the various categories shown in Table 7. It is estimated that the error in determining the composition of the individual grains does not exceed 5% An.

T A B L E 7

QUANTITATIVE PLAGIOCLASE COMPOSITION OF TUFFS AND SANDSTONES (%).

(Twinned plagioclases only)

Unit & Rock type	Oligo-clase	Ande-sine	Labra-dorite
Basal Enon Sandstone (300 grains)	85	15	-
Coerney & Mimosa Fm.: reworked or waterlaid tuffs (100 grains)	77	22	1
Mimosa Fm.: pale red tuff (200 grains)	82	15	3
Coerney Fm.: tuffs (400 grains)	86	13	1

Practically all the andesine grains have a composition close to the An_{30} boundary. The few labradorite crystals noted are most probably accessory constituents derived from the basalt. The average composition of all the grains falls well within the range An_{20}-An_{30}. This corresponds very well with the composition of plagioclases most commonly found in tuffs (Pittman, 1970).

The correspondence in plagioclase content of the Suurberg tuffs and the Basal Enon Sandstone would further suggest that much of the material of the latter unit was derived from the tuffs. Unless labradorite was largely elimi= nated by weathering and transport, it appears as if the basalt contributed practically nothing to the Basal Enon Sandstone, and this in spite of the fact that some of the samples analysed were taken almost immediately above the basalt-Enon contact.

2. Heavy Minerals

a) Introduction

A survey of the literature on heavy minerals shows that there is some disagreement among authors on the resistance of various heavy minerals during the processes of production (weathering), transport, and diagenesis. This is rather unfortunate because sound geological deductions from heavy minerals are dependent on a knowledge of their relative stability. Nevertheless, it is generally accepted that the presence of certain groups of heavy minerals in sediments can be fairly diagnostic of specific source rocks.

In contrast to normal clastics, weathering of the parent rock and stream abrasion play no part in the elimination of heavy minerals of pyroclastic rocks. It is logical then to expect that the heavy minerals of tuffs should bear a direct relationship to the parent magma and physical conditions at the time of eruption. The only exceptions would be in cases where significant amounts of accidental or accessory materials are present, or where postdepositional changes have mate= rially affected the materials.

Published information on the heavy mineral content of pyroclastics appears to be almost non-existent. Although the aim of the present heavy mineral in= vestigation is mainly to obtain additional comparative information on the Suur= berg and Uitenhage Groups, it should at the same time provide some basic informa= tion on the composition of the much neglected pyroclastic rocks.

b) Technique

Heavy minerals were separated with a superpanner from representative samples. A superpanner has some distinct advantages over conventional heavy liquid separations. A more representative crop of all heavy minerals present can be obtained because larger quantities and practically all grain sizes can be treated. It is also quicker and less elaborate and with some skill a com= paratively clean separation between light and heavy fractions can be achieved. Micas and other similar minerals are notable exceptions.

Between 300 and 500 grains per sample were counted by means of a mechanical stage. Minerals of the mica and chlorite groups as well as secondary iron oxides were disregarded.

T A B L E 8

QUANTITATIVE HEAVY MINERAL CONTENT IN PER CENT

SAMPLE	ZIRCON euhedral sharp corners	ZIRCON euhedral slightly rounded	ZIRCON rounded	TOURMALINE brown	TOURMALINE green	TOURMALINE blue/pink	RUTILE red	RUTILE brown	RUTILE yellow	GARNET colourless	GARNET pink	APATITE	EPIDOTE	SPHENE	AMPHIBOLE	MAGNETITE & ILMENITE	UNIT
22	0,6	11,9	46,9	7,1	3,0	-	1,8	9,6	1,8	0,6	-	1,8	1,2	-	0,6	13,1	Sst (ENON FORMATION)
21	0,3	9,4	22,7	2,6	1,6	-	1,0	2,3	0,3	12,4	1,3	0,7	1,6	12,0	0,3	31,5	Basal Enon Sand= stone (ENON FORMATION)
20	0,3	8,2	22,1	0,3	0,6	-	4,0	1,5	1,2	6,7	1,5	0,9	1,2	2,4	2,4	46,7	
19	0,3	9,3	17,8	1,6	1,4	0,2	-	0,8	0,3	10,9	1,6	0,3	8,7	8,2	-	38,6	
18	0,9	7,4	22,5	2,7	1,2	-	2,2	1,8	2,1	15,8	0,3	0,6	7,4	0,6	0,9	33,6	
mean	0,4	8,6	21,3	1,8	1,2	0,1	1,8	1,6	1,0	11,4	1,2	0,6	4,7	5,8	0,9	37,6	
17	-	7,7	22,0	1,2	0,6	-	1,8	0,3	2,7	15,5	0,9	0,6	2,4	1,5	-	42,8	Tuffs (SUURBERG GROUP)
16	0,2	6,2	17,3	1,0	0,4	0,4	0,4	0,7	0,4	6,6	0,4	1,0	3,3	1,9	13,6	46,2	
15	0,7	11,4	35,8	-	-	-	-	-	-	0,7	-	1,4	2,1	1,4	-	46,5	
14	2,9	19,8	35,4	1,2	0,6	-	1,7	-	-	0,6	-	1,2	2,9	1,2	2,9	29,6	
13	2,1	25,4	14,8	2,1	0,9	-	0,4	1,3	0,4	1,7	0,4	31,4	3,8	6,4	-	8,9	
12	1,0	6,8	17,4	0,5	-	-	0,5	-	-	4,5	0,5	4,8	0,8	5,5	24,9	32,8	
11	1,1	6,2	18,7	1,1	1,1	0,2	1,3	2,6	0,5	12,9	0,2	0,8	3,6	3,4	1,8	44,5	
10	0,5	7,6	14,5	2,8	1,3	-	1,3	1,0	1,0	10,4	0,3	2,0	3,3	3,6	1,5	46,1	
9	0,2	6,0	13,2	1,6	0,8	0,4	0,4	1,0	1,0	14,0	1,0	9,1	1,6	1,0	0,4	48,3	
8	-	4,7	19,0	1,1	0,2	0,2	-	0,9	1,1	11,6	1,6	0,4	7,8	2,7	-	48,7	
7	-	3,1	16,5	1,4	0,8	-	1,2	0,2	1,6	10,6	1,2	5,1	7,7	3,3	0,4	48,9	
6	0,3	5,5	27,0	2,9	0,6	-	3,6	0,5	3,2	7,5	0,3	0,9	3,9	4,2	0,9	38,3	
5	-	12,9	19,5	2,0	1,0	0,3	0,3	1,3	0,3	11,0	1,0	3,6	4,9	4,9	0,3	36,7	
4	0,8	6,2	19,6	3,5	0,3	-	1,1	0,8	1,4	9,9	0,5	1,1	8,0	1,9	-	44,9	
3	0,7	5,1	17,3	2,1	1,4	0,7	0,2	0,7	0,7	1,8	0,2	2,8	10,4	2,8	-	53,1	
2	1,4	11,1	32,0	1,8	3,1	0,9	2,2	-	3,5	15,6	1,8	0,9	1,4	3,5	-	20,8	
mean	0,7	8,6	20,9	2,2	1,2	0,2	1,1	1,5	1,2	9,3	0,8	4,2	4,6	3,8	2,1	37,6	
1	-	9,2	42,2	2,0	0,5	-	7,2	6,1	2,0	3,6	-	1,0	2,6	-	-	23,6	Sst. (SUURBERG GROUP)

SAMPLE

22 - Paarden Laaghte (B) : sandstone within Enon Conglomerate.

21 - Erenkrons Poort (A)
20 - Beyers Vley Outspan (B)
19 - Erenkrons Poort (A)
18 - Rhinoster Hoek No. 2 (B)
: Basal Enon Sandstone.

17 - Rhinoster Hoek No. 2 (B) : tubular tuff overlying Mimosa basalt.

16 - Slag Boom (E) : water-laid tuff interbedded with Mimosa basalt.

15 - Enon Mission Station (E)
14 - Beyers Vley Outspan (B)
: pale red welded tuff interbedded with Mimosa basalt.

13 - Woodbury (G)
12 - Beyers Vley Outspan (B)
: pale whitish tuff, Coerney Formation.

11 - Beyers Vley Outspan (B)
10 - Springbok Vlackte No. 6 (B)
9 - Rhinoster Hoek No. 2 (B)
8 - Slag Boom (E)
7 & 6 - Drie Kuilen (A)
5 & 4 - Erenkrons Poort (A)
3 - The Retreat No. 1 (G)
2 - Blaauwbosch Kuil (C)
: orange/pink tuff, Coerney Formation.

1 - Tover Klip (C) : sandstone from Slagboom Formation.

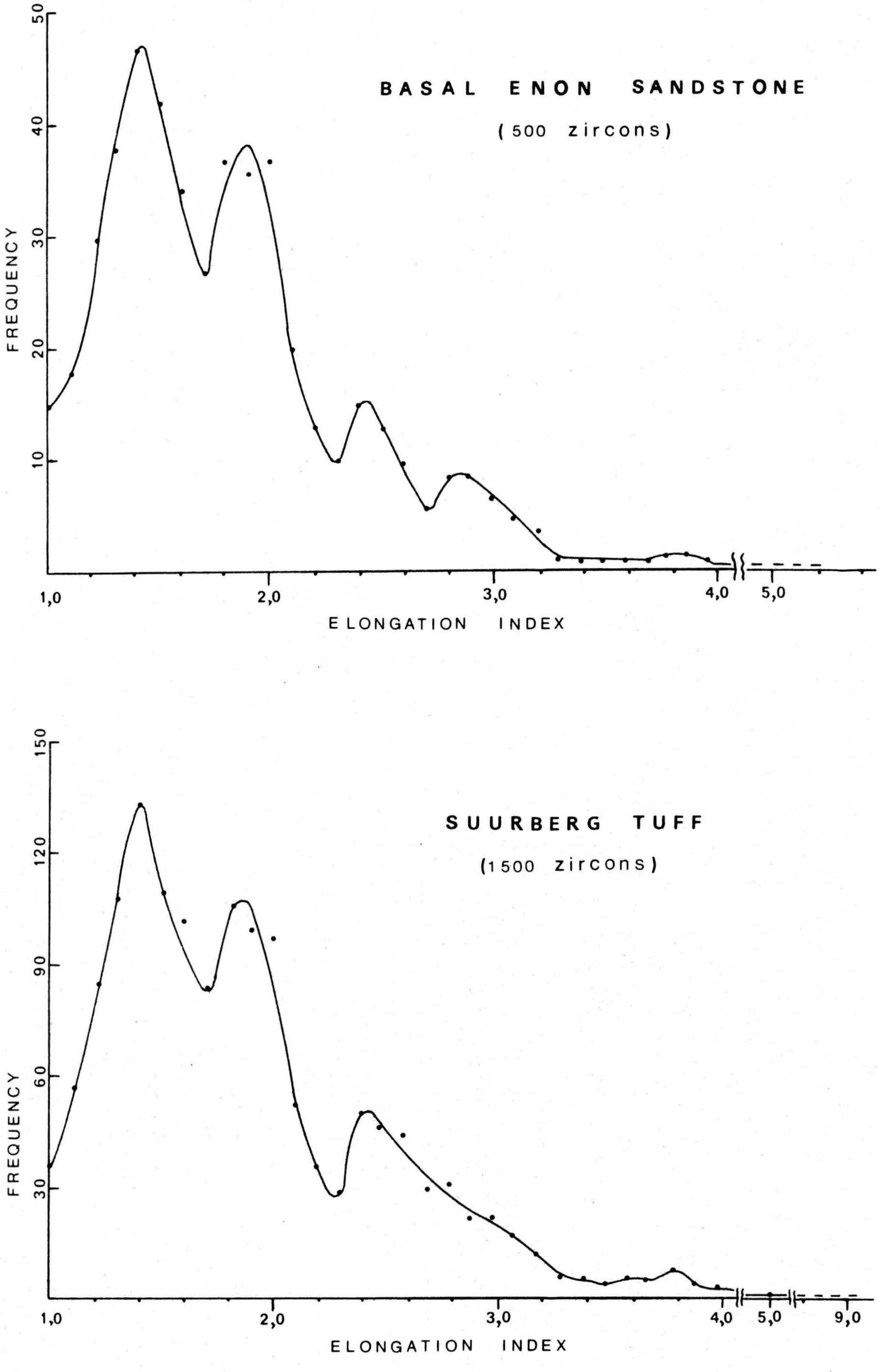


FIG. II. COMPARATIVE ZIRCON ELONGATION INDEX DIAGRAMS

c) Description and significance of species

The results of the heavy mineral count are presented in Table 8.

The zircons are colourless or feebly coloured and pleochroic. Of 500 zircon grains counted in the Basal Enon Sandstone 58% were coloured and the same applied to 32% of 1500 grains from the Suurberg tuffs. Both zoned and unzoned varieties occur.

Elongation index diagrams of the zircons from the above-mentioned two units are presented in Figure 11. The similarity of the diagrams are striking, with major peaks at length : breadth ratios of 1,4 and 1,9 , and secondary peaks between 2,4 and 2,9.

Zircon is generally regarded to be associated with acid igneous rocks when euhedral, or with reworked sediments when well rounded. However, it is known that rounding of zircons can take place by magmatic resorption, especially in effusive igneous rocks (Deer et al., 1966, p. 16).

Most tourmalines are prismatic and well rounded grains are scarce. Tour=maline is usually associated with rocks of acid affinity (Deer et al., 1966, p. 95).

Only well rounded ellipsoidal grains of rutile are present. This mineral apparently enjoys such a wide and general distribution that it is not of much diag=nostic value.

Garnet grains are comparatively large and although both rounded and euhedral grains with dodecahedral faces are present, irregular grains, bounded by conchoidal fracture surfaces, predominate. Colourless grains outnumber pink ones by about 10 : 1 in both the Suurberg tuffs and the Basal Enon Sandstone. The refractive index ($n \simeq 1{,}800$) falls in the range for almandine and spessartite. Although garnets are usually associated with metamorphic rocks they also occur in acid volcanic rocks (Deer et al., 1966, p. 29).

Most apatite grains are prismatic with rounded terminations. According to its refractive index, $\omega = 1{,}640$ ($\pm 0{,}005$), and very low birefringence it is a fluor-apatite (Deer et al., 1966, p. 507). The distribution of apatite is erratic in the tuffs and in a few samples it is a very prominent heavy mineral. This sporadic occurrence is usually an indication of a volcanic source (Folk, 1968, p. 97).

With epidote in Table 8 clinozoisite is also included. The latter occurs as colourless prismatic crystals while the variety epidote, which predominates, is usually moderately rounded, faintly coloured and pleochroic.

Sphene appears to be a characteristic heavy mineral of the Coerney tuffs and the overlying Basal Enon Sandstone. It is easily identified by its high relief, golden-yellow colour with slight pleochroism, and the anomalous extinction which produces an intense blue interference colour. The crystal faces are well de=veloped and smooth. Sphene is usually associated with acid to intermediate igneous rocks (Deer et al., 1966, p. 20).

Under the heading "Amphibole" in Table 8 are grouped mainly hornblende and basaltic hornblende. A few doubtful, but apparently unstable species have also been included. Basaltic hornblende is often thought to have been produced in nature from normal hornblende at temperatures above 800°C. (Deer et al., 1966, p. 176).

Magnetite and ilmenite are not difficult to recognize in reflected light, but it is less feasible to distinguish between the two except for occasional euhedral grains. Investigation with a magnet showed that magnetite was dominant.

d) Some deductions from the heavy minerals

The more important features of the heavy mineral assemblage are the following:

i) The crop as a whole is indicative of an acid igneous source (Pettijohn, 1957, p. 513).

ii) Fair amounts of metastable species like apatite, epidote, hornblende, and sphene are present (Pettijohn, 1957, p. 504).

iii) Minerals like sphene, epidote, and basaltic hornblende could have been deuteric in origin.

iv) Garnet, epidote, hornblende, sphene, and even coloured varieties of zircon are very scarce or absent in rocks of the Cape Supergroup which largely surround the Algoa Basin (De Villiers & Wardaugh, 1962; Loock, 1967; Theron, 1970; etc.). It should be noted, however, that garnet is a prominent heavy mineral of the Dwyka Group (Loock, 1967). Even so, it would appear as if it did not contribute materially to the Suurberg Group and the lowermost Enon. In this connection sample no. 2 in Table 8 is significant, because it comes from a locality where contribu= tion from Dwyka sources would have been extremely unlikely.

From the foregoing it is concluded that practically all the heavy minerals of the Suurberg tuffs were derived from an acid parent magma, though some might have resulted from deuteric processes associated with the volcanism. No note= worthy amounts of accidental or externally derived materials appear to be present.

Probably the most significant fact which emerges from Table 8 and Figure 11 is the remarkable correspondence between the heavy mineral assemblages of the Suurberg tuffs and the Basal Enon Sandstone, which normally would indicate a common source type. In this connection two possibilities arise.

Firstly, the Suurberg tuffs could have acted as source for a large portion of the materials incorporated into the Basal Enon Sandstone. This statement seems to be supported by the many tuff fragments present in the basal Enon of the southern and western Panhandle. If most of the heavy minerals were derived from the tuffs it is further required that their deposition should have taken place before prolonged weathering or transport could have eliminated unstable species. It is for instance known that in situ rock weathering can destroy even garnet (Dryden and Dryden, 1946). Ideally then unconsolidated or only partly consolidated ash and only limited local dispersal would be necessary. It is also noteworthy that many rock fragments from the well indurated Cape Supergroup are present in the Basal Enon Sandstone but apparently very few heavy minerals have been derived from this source. This provides additional evidence for limited breakdown and dispersal of materials.

It is further of interest that no basalt fragments or minerals typical of basalts were noted in the basal Enon sandstones. This might be interpreted in one of two ways, namely, that either basalt material was not readily available, or that it was destroyed before deposition. This last point would be contradictory with the indications mentioned in the preceding paragraph. The first mentioned possibility also appears improbable as a glance at Figure 2 would show — except for the area east of Rhinoster Hoek no. 2 (B) the Basal Enon Sandstone is practically everywhere underlain by basalt. An obvious solution for the problem would be the reasonable assumption that, at the commencement of Enon times, the southern Panhandle depository was filled with largely unconsolidated volcanic ash which occurred higher up, outside the confines of the basalt filled troughs, rather than with materials from the low lying and crystalline basalt itself.

Alternatively, the volcanism could still have been active in certain parts of the area while the Basal Enon Sandstone was being deposited. In this case pyroclastic material could have made a major mineralogical contribution to the finer Enon sediments. An analogous situation of simultaneous deposition of volcanic and detrital materials probably existed in Cave Sandstone times (Late Triassic to Early Jurassic). Volcanism was then already active (Botha & Theron, 1967), and De Villiers & Wardaugh (1962) mention the sudden prominence of epidote in heavy concentrates of the Cave Sandstone, in contrast to its usual absence in rocks of the Cape and Karoo Supergroups.

The tuff which overlies the basalt on Rhinoster Hoek no. 2 (B) might indicate that the volcanism in certain parts of the Panhandle was concluded with local tephra ejections and not with basalt extrusions as appears to be the case else= where in the Algoa Basin (see Figure 2).

All the foregoing explanations assume only a small time lapse, or even some overlapping, between the Suurberg volcanism and the beginning of Enon deposition.

D. PETROCHEMISTRY

When the present study started there was only one published chemical analysis of the Suurberg lava (Haughton and Rogers, 1924, p. 245). Because of the high proportion of (glassy) matrix in the basalts and tuffs further chemical analyses were deemed necessary to supplement the modal analyses. At the same time this would enable comparison with similar rocks from elsewhere.

Six samples were submitted through the Geological Survey to the National Institute for Metallurgy for whole rock analyses. Except for FeO, which was determined by a volumetric procedure, and H_2O (gravimetric procedures), values for all other major elements were obtained by X-ray fluorescence methods. The results are shown in Table 9.

TABLE 9

CHEMICAL ANALYSES OF SUURBERG ROCKS

Sample	1	2	3	4	5	6	7
SiO_2	69,69	67,56	75,60	80,53	50,32	50,20	50,75
TiO_2	0,27	0,19	0,31	0,24	0,88	1,09	1,15
Al_2O_3	12,60	12,74	11,67	8,66	15,19	14,47	13,8
Fe_2O_3	0,91	0,44	1,34	0,99	3,03	7,26	4,65
FeO	0,23	0,28	0,25	0,14	6,87	3,98	6,2
MnO	0,11	0,06	0,06	0,04	0,22	0,20	0,1
MgO	0,99	0,98	0,36	0,58	7,18	5,32	7,1
CaO	2,55	2,11	1,26	1,42	11,03	8,29	8,9
Na_2O	0,45	1,14	2,24	0,87	2,11	4,19	2,85
K_2O	3,22	3,11	4,26	2,88	0,74	0,98	0,85
P_2O_5	0,06	0,03	0,55	0,06	0,20	0,22	0,2
H_2O^+	5,45	8,22	1,58	2,58	1,23	3,00	2,35
H_2O^-	3,24	2,30	0,48	0,52	1,17	0,90	1,85
Totals	99,77	99,16	99,96	99,51	100,17	100,10	100,75

Analysts (1-6): General Superintendence Co. (Pty) Ltd.

VOLUMETRIC MODES

	1	2	3	4	5	6	7
Quartz	5,8	Q,3	8,2	7,3	-	-	
K-feldspar	7,5	4,6	11,6	7,7	-	-	
Plagioclase	4,8	1,9	7,6	7,9	47,1	35,7	p
C-Pyroxene	-	-	-	-	41,4	4,0	p
Biotite	0,8	0,4	-	1,4	-	-	
Ore	0,3	0,1	0,7	0,6	2,5	4,9	p
Olivine	-	-	-	-	-	-	p
Other	0,7	0,6	1,6	1,1	9,0	-	
Groundmass	80,1	92,1	70,3	74,0	-	51,1	p
Amygdales	-	-	-	-	-	4,3+	p
Totals	100,0	100,0	100,0	100,0	100,0	100,0	-

+ mainly celadonite p = present but not determined

1 - Pale orange/pink tuff, Coerney Fm. ; Springbok Vlackte no. 6 (A)
2 - Pale whitish tuff, Coerney Fm. ; Rhinoster Hoek no. 2 (B)
3 - Pale red tuff, Mimosa Fm. ; De Vlei (A)
4 - Pale red tuff dyke, Mimosa Fm. ; Beyers Vley Outspan (B)
5 - Dolerite, intrusive in Coerney Fm. ; Slag Boom (E)
6 - Basalt, Mimosa Fm. ; Drie Kuilen (A)
7 - Basalt, Mimosa Fm. ; De Vlei (A). (Haughton & Rogers, 1924)

A striking feature of the analyses of the Suurberg rocks is the high water content, which is believed to indicate the presence of hydrous minerals like zeolites and celadonite, and glass. Zeolites probably occur optically irresolvable in the matrix of some tuffs. In the basalt the zeolites and celadonite of the vesicles are usually minutely intergrown so that it becomes an almost impossible task to calculate and apply "corrections" to take care of the secondary minerals.

The very high proportion of ferric iron to total iron in the basalts is probably largely due to oxidation during subaerial extrusion, a process which did not affect the intrusive dolerite to the same extent, or to alteration during zeolitization of the lava pile (Cox & Hornung, 1966, p. 1423). For norm calculations and comparative purposes the effect of such accidental oxidation should be removed.

Thornton and Tuttle (1960) show that most of the 5 000 superior analyses of basaltic and other igneous rocks have a remarkably consistent Fe_2O_3 value of approximately 2,5% by weight. Other authors have introduced various correction factors to offset the effects of secondary oxidation of the iron in basalts. The present author has reduced the Fe_2O_3 value for the anhydrous Suurberg basalts to 3,0%, with a corresponding increase in FeO, to approach the oxidation state of the Suurberg dolerite more closely.

Table 10 compares the analyses of Suurberg rocks, calculated on an anhydrous basis, with those of similar rocks from elsewhere. Basalts from Lesotho and rhyolitic rocks from the Lebombo Range are included because they are geographically the nearest equivalents of the Suurberg volcanics. The rhyolitic rocks from the Andes are also represented in Table 10 because, as far as tectonic setting and silica content are concerned, they correspond closer to the Suurberg acidic rocks than those of the Lebombo.

Experimental studies by various workers have provided evidence which shows that fractional crystallization of complex magmas produces liquids which move toward and, for all practical purposes, eventually reach the system SiO_2 - $NaAlSiO_4$ - $KAlSiO_4$ (petrogeny's residua system). Thornton and Tuttle (1960) introduced the term Differentiation Index, a value which gives a quantitative indication of how far a given magma has travelled towards this goal, thus providing a simple measure of the "basicity" of an igneous rock.

In Figure 12a (after Thornton & Tuttle, 1960, p. 675) Silica versus Differentiation Index is plotted for the analyses from Table 10. Figure 12a also illustrates the silica content of normative Orthoclase, Albite and Anorthite. Oversaturated rocks will generally fall above the line connecting Or and An whereas saturated and undersaturated ones will fall below this line. The same diagram is further subdivided into acidic, intermediate, basic and ultrabasic fields.

The chemical analyses clearly confirm that the Suurberg rocks belong to two distinct categories, acidic and basic, with no intermediate types present.

TABLE 10

ANALYSES OF DIFFERENT ROCK TYPES (ANHYDROUS BASIS)

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	76,52	76,22	77,22	83,53	51,47	52,18	52,55	51,8	52,7	51,6	69,34	76,8
TiO_2	0,29	0,21	0,32	0,25	0,90	1,13	1,2	1,13	1,16	1,6	0,81	0,14
Al_2O_3	13,83	14,37	11,92	8,98	15,54	15,04	14,3	14,8	15,4	16,0	13,14	12,43
Fe_2O_3	1,00	0,50	1,37	1,03	3,10	3,00	3,0	3,92	1,38	2,9	4,77	0,8
FeO	0,25	0,32	0,26	0,15	7,03	8,69	8,2	7,26	9,35	8,1	1,03	0,13
MnO	0,12	0,07	0,06	0,04	0,22	0,20	0,1	0,17	0,22	0,2	0,09	0,04
MgO	1,09	1,11	0,37	0,59	7,34	5,53	7,35	7,1	6,6	6,2	0,32	0,23
CaO	2,80	2,38	1,29	1,47	11,28	8,62	9,2	10,57	9,96	10,0	1,71	0,84
Na_2O	0,49	1,29	2,29	0,90	2,16	4,36	2,95	2,40	2,22	2,4	3,71	3,65
K_2O	3,54	3,51	4,35	2,99	0,76	1,02	0,9	0,74	0,87	0,7	4,89	4,87
P_2O_5	0,07	0,03	0,56	0,07	0,20	0,22	0,2	0,13	0,16	0,2	0,18	0,02
Totals	100,00	100,01	100,01	100,00	100,00	99,99	99,95	100,2	100,02	99,9	99,99	99,85
* σ	0,5	0,7	1,3	0,3	1,1	3,2	1,0	1,1	1,0	1,1	2,8	2,1

C. I. P. W. NORMS

		1	2	3	4	5	6	7	8	9	10	11	12
Q		52,7	48,6	45,6	63,1	2,45		1,54	3,45	3,19	4,3	25,95	35,48
Or		20,9	20,7	25,7	17,6	4,49	6,01	5,34	4,40	5,12	4,1	28,88	28,77
Ab		4,1	10,9	19,3	7,6	18,27	36,86	24,96	20,29	18,77	20,3	31,40	30,04
An		13,4	11,6	2,8	6,8	30,46	18,47	23,14	27,43	29,50	30,8	4,76	3,59
C		4,2	4,2	2,4	1,8								
Di	Wo					10,05	8,68	8,82	10,12	7,92	7,2	0,92	0,21
	En					6,28	4,97	4,26	6,38	4,40	4,3	0,79	0,18
	Fs					3,17	4,33	3,11	3,11	3,22	2,6		
Hy	En	2,7	2,8	0,9	1,5	12,10	1,44	14,05	11,25	12,03	11,2		0,39
	Fs		0,1			6,07	1,25	7,67	5,43	11,31	6,9		
Ol	Fo						5,16						
	Fa						4,95						
Ap		0,2	0,1	1,3	0,2	0,47	0,50	0,47	0,30	0,37	0,5	0,44	0,03
Mt		0,3	0,7	0,1		4,50	4,35	4,35	5,68	1,99	4,2	0,67	0,04
Il		0,5	0,4	0,6	0,4	1,73	2,14	2,28	2,13	2,17	3,8	1,92	0,33
Hm		0,8		1,3	1,0							4,31	0,77
Totals		99,8	100,1	100,0	100,0	100,04	99,11	100,99	99,97	99,99	100,2	100,04	99.83
+D.I.		77,7	80,2	90,6	88,3	25,2	42,9	31,8	28,1	27,1	28,7	90,2	94,3

+Differentiation Index (Thornton & Tuttle, 1960) = Q + Or + Ab

*Suite index (Rittmann, 1962) = $\frac{(Na_2O + K_2O)^2}{SiO_2 - 43}$ (weight % in analysis)

1–7 Suurberg rocks, recalculated from Table 9.
8 Average of 21 Lesotho basalts (Cox et al., 1967, p. 1462).
9 Average of 44 Karoo dolerites (Cox et al., 1967, p. 1462, from data of Walker & Poldervaart, 1949).
10 Average of 1228 tholeiitic basalts and dolerites (Manson, 1967, p. 227).
11 "Rhyolitic Series", average of 23 analyses from Lebombo Mountains (de Assuncao et al., 1962; quoted by Stratten, 1965).
12 Average of 9 alkali rhyolitic ash-flow tuffs from the Andes, Chile (Zeil & Pichler, 1967).

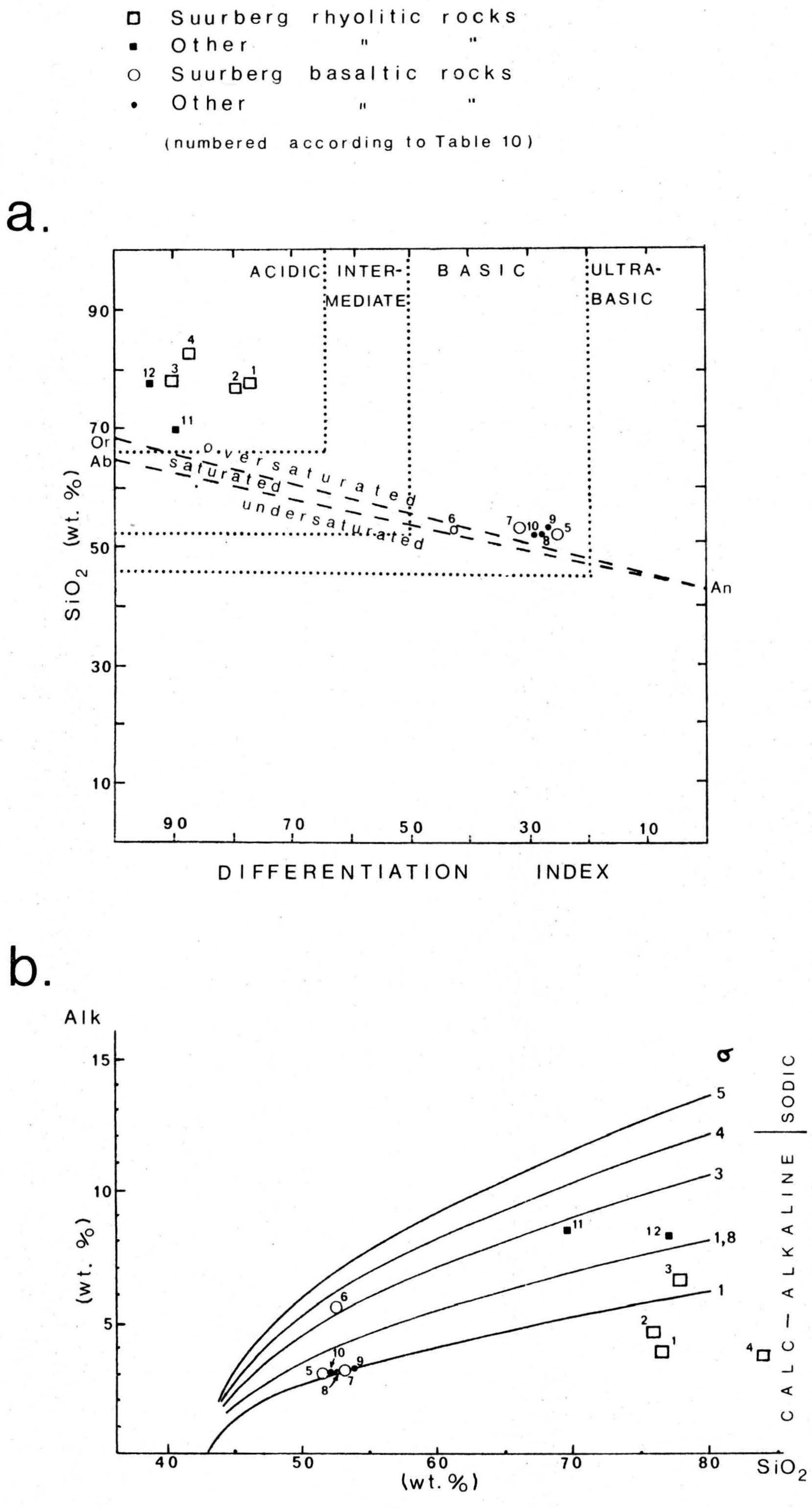


FIG. 12. CHEMICAL VARIATION DIAGRAMS

(SEE TEXT FOR DETAILS)

The Suurberg basic rocks fall into the class Tholeiitic Basalt (Dolerite) of Yoder and Tilley (1962) due to the presence of normative quartz and hypersthene. Except for anomalously high Na_2O (possibly due to zeolites in vesicles) the Suurberg basalts also have a close chemical correspondence with their Karoo counterparts. Furthermore, when plotted on the variation diagrams of Cox et al. (1967), which are not reproduced here, the Suurberg basalts invariably fall in the field of their Southern Basalt Province.

In the analyses of the Suurberg tuffs the CaO values appear to be rather high and the Na_2O too low. This results in abnormal feldspar proportions in the norm. The Na_2O deficiency probably also led to the appearance of normative corundum, which suggests leaching of Na_2O. These and other anomalies are not altogether unexpected for tuffs which, on account of their original porosity, are more subject to allochemical changes than crystalline rocks. Even so, it is believed that these secondary changes have not obliterated the overall picture. The chemical analyses have confirmed the acidic nature of the tuffs although the silica content is significantly higher than formerly deduced from a study of their crystal fraction alone. This appears to be normal (p. 28) and was already suggested by the refractive indices of some unaltered glass shards.

The relation between silica and alkalis for the analyses listed in Table 10 are shown in Figure 12 b. The paraboloidal curves represent constant σ- values (Rittmann, 1962, p. 110) which serve to determine the suite character. The calc-alkaline ("Pacific") nature of the Suurberg basalts and tuffs is evident, although some allowance should be made for the obvious Na_2O-deficiency of the latter.

Many igneous rock types are believed to have links with specific tectonic environments and many chemical variation diagrams have been drawn up for this purpose. In these diagrams alkalis usually play an important role so that they would be of little use for the Suurberg rocks with their uncertain Na_2O values. The TiO_2 content of the Suurberg rocks is probably a more reliable value to use for comparative purposes.

Figure 13 below shows two pronounced maxima for TiO_2 content of lavas from active volcanoes. The first maximum corresponds to the volcanites of orogenic belts, the second is due to basaltic rocks from cratonic zones, while a third minor peak corresponds to the alkali basalts and tephrites (Gottini, 1968, p. 933). The overall trend is that TiO_2 is higher for igneous rocks from cratonic areas than for those from orogenes.

The rhyolitic rocks from the known folded regions (analyses 1-4 and 12 in Table 10) have consistently low TiO_2 values. Higher values are found in the Lebombo rhyolitic rocks which occur along the eastern margin of the Kaapvaal Craton, parallel to the strike of the Mozambique Mobile Belt (Anhaeusser et al., 1968, p. 3). At the same time the last mentioned rocks have lower SiO_2 and much higher total iron values. The TiO_2 content of Suurberg basaltic rocks is only slightly higher than the first peak in Figure 13, but this is also appli=

cable for most basalts from the Southern Province of Cox et al. (1967). However,

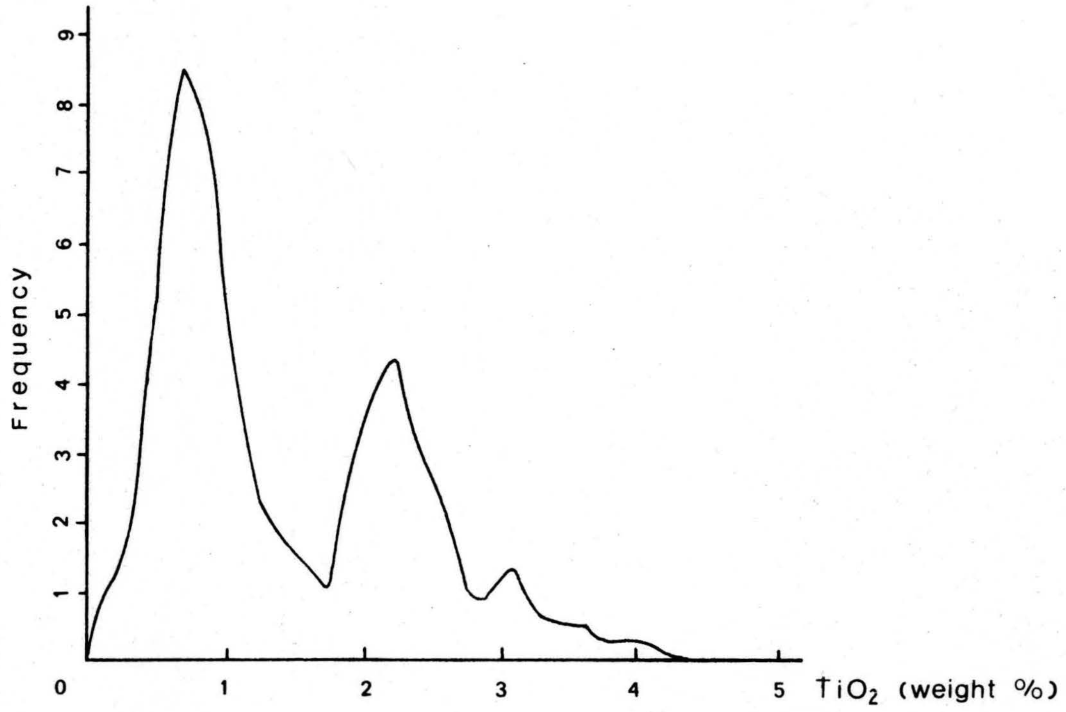


Fig. 13 Frequency of the TiO_2 content in the lavas of active volcanoes (Gottini, 1968).

it is notable that for Swaziland basalts, which are associated with the Lebombo rhyolitic rocks, TiO_2 increases to 1,65% (op. cit., p. 1462). For basalts of the Northern Province the TiO_2 values are still higher and correspond with those for alkaline rocks from cratonic regions.

These trends will be further analysed in Chapter V.

IV STRUCTURAL FEATURES

In this study attention was paid to the nature of contacts, the attitude of beds, and the presence of joints and faults. The main objective was to determine the correct stratigraphic interrelationship of the various units and to gather some data with bearing on the earlier tectonic history of the Algoa Basin.

A. NATURE OF THE CONTACTS

Reference has already been made to this aspect in the course of field descriptions (Chapter II). To summarize:

i) Rocks of the Cape-Karoo sequence form the outer rim of the Algoa Basin and are separated from the overlying Suurberg and Enon by a pronounced angular unconformity. This unconformity is well exposed in several localities (pp. 9, 21).

ii) The rock units of the Suurberg Group form a practically continuous stratigraphic sequence with gradational or conformable boundaries. This is for instance well exposed in a road cutting on Slag Boom (E).

iii) The Enon apparently has a conformable relationship with the underlying Mimosa Formation, the contact being a sedimentary one. (p. 21). At the western and possibly also the eastern extremities of the volcanic rock outcrops the Enon is separated from the Coerney tuffs by an erosional break, which appears to be of minor proportions.

The statement made under i) above differs from published maps and descriptions which show a tectonic contact between the Paleozoic and Mesozoic rocks of this area. Several faults of post-Enon age have been postulated along the margin of the Algoa Basin, of which the most important, the "Zuurberg Fault", has become well established in the literature.

It is notable that the earliest descriptions mention only indirect evidence for the existence of the "Zuurberg Fault". Rogers (1905, p. 190) first postulated a post-Enon normal fault, with downthrow to the south, along the southern flank of the Suurberg range in order to explain the great variation in width of the Enon conglomerate in this vicinity. The same author added: "The increased, but varying dip of the conglomerates towards their northern boundary supports this supposition which is in accord with the observations of Mr. Schwarz in the Willowmore and Uniondale Divisions, that the outliers of the Uitenhage beds there are faulted down against the older rocks to the north of them." Rogers (1906, p. 18) further wrote: "The fault explains the remarkable straight northern boundary of the (Enon) formation and the apparently very steep angle at which the Witteberg quartzites disappear under it." He also explained the presence of the volcanics in that region by suggesting that they had risen along the line of faulting.

Later, when Haughton and Rogers (1924) revoked some of Rogers' original conclusions, and correlated the volcanics with the Stormberg extrusives, the "Zuurberg Fault" was retained and extended westwards.

The present author cannot support the existence of the "Zuurberg Fault" on several grounds:

i) Rapid thickness variations are to be expected with torrential conti= nental deposits so that it appears unnecessary to resort to faulting to explain the "great variation in width" in the case of the Enon conglomerate. In part this "variation in width" is probably also more apparent than real because the upper boundary of the Enon conglomerate is poorly defined and, more often than not, covered by superficial de= posits.

ii) The "increased dip of the Enon conglomerates towards their northern boundary" is not unusual because dips normally increase from the centre towards the flanks of a depositional basin. This aspect will be treated further in a later section.

iii) Cleavage planes in the tuffs, mistaken for bedding, probably contri= buted to the wrong conclusion (p. 14).

iv) The more detailed mapping shows that the northern termination of the Enon Conglomerate and the Suurberg volcanics is highly irregular with evidence of control by paleotopography rather than post-Enon faulting (see maps 1 and 2).

v) The actual discordance between the Paleozoic "floor" and the Mesozoic rocks along the northern margin of the Algoa Basin is observable at several localities (see p. 21). The same cannot be said of the "Zuur= berg Fault".

In a like manner the existence of other faults, previously shown to have de= lineated the western and southern Kirkwood Panhandle, can be discredited.

B. FAULTS

It was shown above that the previously suggested faulted contacts along the periphery of the Algoa Basin could not be verified in the field. In the Bush= mans River area Haughton and Rogers (1924, p. 244) also postulated a series of three wrench faults with approximately north-south strike. This area is ex= tensively covered by Cainozoic deposits so that these faults could only have been inferred. On the other hand, their supposed effect of displacing the Witteberg progressively southward and thereby causing discontinuous outcrops of quartzites and volcanics, could equally well have been explained by a series of pitching anticlines and synclines. This being the case elsewhere in the same area, it seems unnecessary to retain these wrench faults for the present.

There is, however, field evidence for the existence of pre-Enon strike

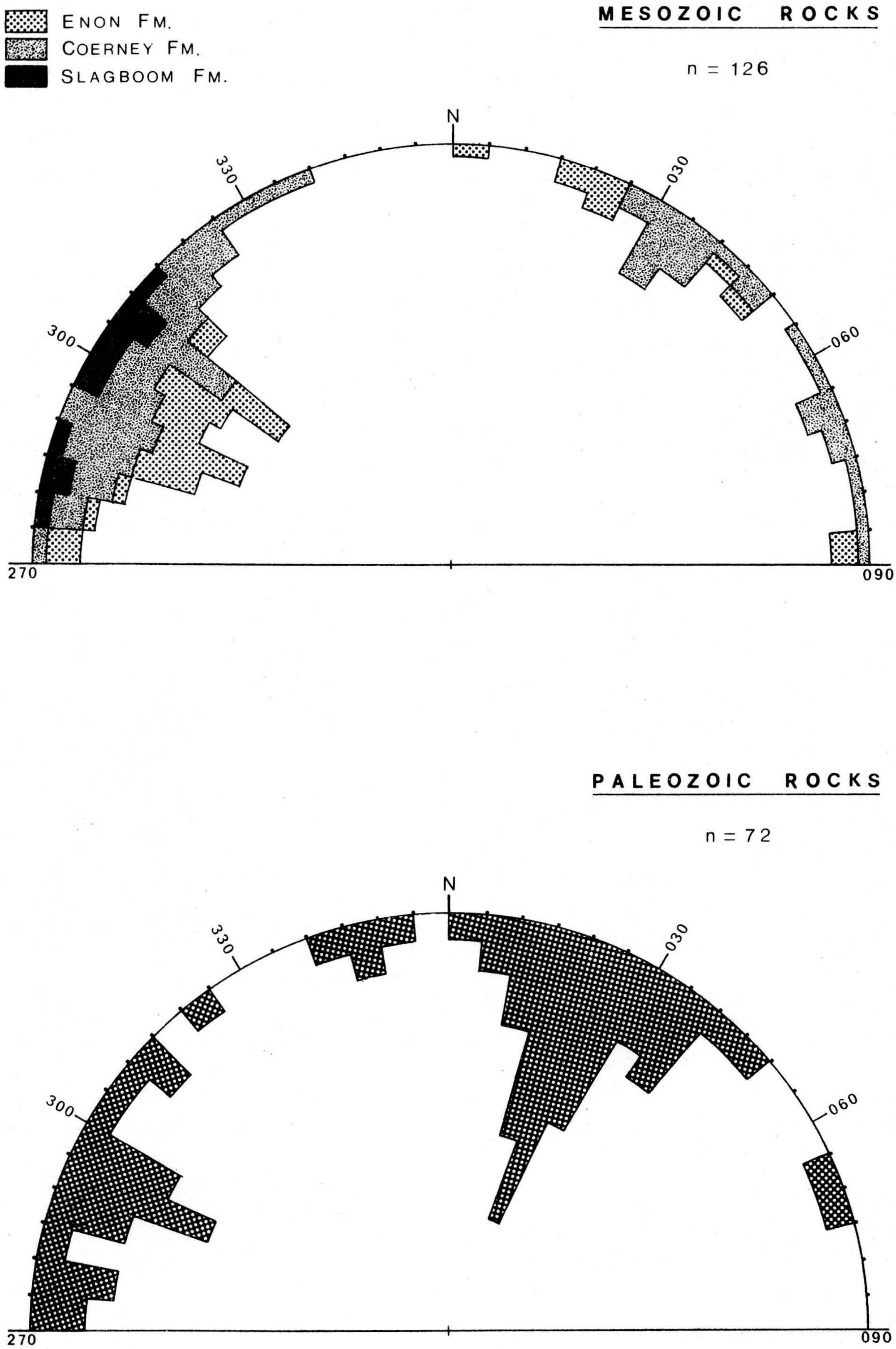


FIGURE 14

STRIKE-FREQUENCY OF SETS OF VERTICAL TO SUB-VERTICAL JOINTS AND FRACTURES faults in the Cape-Karoo floor of the Algoa Basin. The most important surface example of such a fault was found in the Witteberg rocks on Botha's Kraal (D) (Fig. 9). Clear evidence of faulting is provided by the major break in the normal stratigraphic sequence, as well as the presence of intensely brecciated quartzite. The fault is responsible for a vertical displacement in the order of 300 metres. It appears to be a normal fault with downthrow to the south. The attitude of the fault plane is not directly observable but it was taken to be dipping steeply, as is usually the case with normal faults (Price, 1966, p. 59).

The Botha's Kraal Fault strikes E.S.E., parallel to the trend of the main fold axes of the Paleozoic rocks. The magnitude of this fault decreases west=wards. Eastwards it passes underneath unaffected Enon beds and its further course could perhaps be inferred from the brecciated nature of the Witteberg rocks along the margin of the basin locally. This brecciation decreases progressively east=wards within 2 kilometres, probably in sympathy with the increasing distance from the inferred course of the fault (see Map 1). About 2 or 3 kilometres west of Kirkwood a protruding nose of Witteberg is intensely brecciated suggesting a possible further extension of the Botha's Kraal Fault.

Small normal faults with displacements of a few metres and less are not unusual in the Mesozoic rocks along the margin of the Algoa Basin. Good examples occur on Beyers Vley Outspan (B) and Slag Boom (E). The downthrow is invariably towards the basin centre so that these small faults evidently reflect adjustment of the layers to a subsiding depository.

C. JOINTS

Strike directions of sets of vertical to near-vertical joints or fractures which occur most commonly in the investigated area are shown in Figure 14.

In the Paleozoic rocks the most prominent strike directions of joints, N.N.E. with a secondary trend developed more or less at right angles to it, ob=viously belong to the system of tension joints associated with the E.S.E. striking axes of the Cape Folds in this area (cf. Theron, 1962, p. 385).

In the Suurberg and Enon rocks the dominant strike direction of joint sets, namely W.N.W., is particularly borne out by the closely spaced parallel frac=tures which occur in the rudites. These fractures are best developed at or near the base of the Mesozoic rocks. They do not deviate around clasts in the conglomerates and breccias and in this respect they have the nature of shear joints (Price, 1966, p. 121), although this is not a reliable criterion in well cemented rocks (Billings, 1954, p. 118). Also, no movement is apparent along the fracture planes. There are no intense post-Enon folds or faults to bear witness to severe compressive stresses or couples that could have produced shear joints. Furthermore, similar fractures occur in isolated quartzite fragments which are embedded in tuffs of quite different competence so that an origin due to normal tensile stresses can also be ruled out.

In the light of the foregoing evidence it would appear as if the pheno= menon can best be explained by the mechanism which produces release fractures (Billings, 1954, p. 96-97). Fractures normal to the greatest principal stress axis may develop in material when load release follows compression under conditions of high confining pressure. In this particular case then it would appear that the compressive forces acted more or less parallel to the floor of the basin in a direction perpendicular to the long axis of the Panhandle-Sundays River trough. At the same time this direction of compression would be virtu= ally parallel to that which originally acted upon the Cape and Karoo rocks to produce the Cape Folding. It seems reasonable therefore to ascribe the com= pression to "adjustment" of the floor rocks during basin development.

In the Coerney tuffs this compression is manifested by a cleavage which commonly obliterates the faint bedding. A weak foliation might have developed, but this could not be positively confirmed under the microscope.

D. THE ATTITUDE OF BEDS

From geological maps of the area it is immediately apparent that there is a significant structural difference between the Mesozoic rocks within the Algoa Basin and those of the Cape-Karoo floor. In the latter steep dips (over 45°) and even inverted beds prevail, while dips in the younger formations are invaria= bly below 45° and usually much less.

The intense folding of the Cape-Karoo sequence no doubt dates back to the Cape Orogeny which was operative mainly from Late Permian to Mid Triassic times (De Villiers, 1944, p. 202). This folding is characterized by tight asymmetrical folds which are commonly overfolded and locally accompanied by small thrusts. Overfolding is usually to the north, but in the Kirkwood-Paterson area overfolding to the south also occurs, which is rather unusual for the Cape Folded Belt (cf. De Villiers, 1944, p. 193). The folding resulted essentially in a series of pitching synclines and anticlines arranged <u>en echelon</u>. The axes of these folds strike between 100° and 110°.

This structure of the "floor" rocks was subsequently accentuated by erosion and exercised an important control on the distribution of the younger rocks. Evidence of this is provided by the outcrop patterns on Drie Kuilen (A), Erenkrons Poort (A), Schiet Nek (D), northwest and northeast of Kirkwood, north of Mimosa siding (F), and on Solomon's Temple (G) - the reader is referred to Map 1.

Locally irregular dip variations can be detected in the Mesozoic rocks when followed across their strike . Exposures are usually too poor or of too limited extent to allow any definite conclusions on whether low intensity open folding and/or differential compaction can be held responsible for this feature. The first-mentioned possibility would normally have required compressive stress, while differential compaction may take place over buried topography. Both factors were operative in the early Algoa Basin (pp. 47, 13). In the Enon conglomerate it is evident that most dip variations are primarily depositional features.

On Beyers Vley Outspan (B) the basal Enon sandstones and shales are highly deformed (Plate 2B). There can be little doubt that these contortions are the result of slumping, indicating gravitational stress which, on a larger scale, could have been responsible for epidermal compressive settling (Van Bemmelen, 1954) in the subsiding basin.

Dips in the Mesozoic rocks are virtually always directed towards the inner basin and decrease progressively from the margin inwards. Table 11 below shows that the highest dips along the margin of the Algoa Basin occur in the Kirkwood-Paterson area, with a tendency to decrease westwards.

TABLE 11

DIPS ALONG THE INNER MARGIN OF THE NORTHERN ALGOA BASIN

	Area	Number of measurements	Range of dip values	Mean (95% probability)
KIRKWOOD-PATERSON	Eastern	43	15°–40°	$27{,}4^\circ$ ($\pm 2{,}6^\circ$)
	Western	44	15°–30°	$21{,}9^\circ$ ($\pm 1{,}4^\circ$)
PANHANDLE	Northern	46	5°–30°	$20{,}8^\circ$ ($\pm 2{,}2^\circ$)
	Southern	56	5°–25°	$13{,}8^\circ$ ($\pm 1{,}2^\circ$)

From the mean dip values it would appear that the Panhandle trough is asymmetrical with the steeper flank on the northern side; also that there is a pronounced deepening of the basin from Kirkwood eastwards. It is further significant to note that the Enon conglomerates show their best development in this last-mentioned area, while they are very poorly represented at the base of the Enon Formation in the southern Panhandle where low dips prevail.

The distribution pattern of the basalt north of Mimosa (Map 2) clearly indicates that the basalt flows were restricted to the old valley floors and consequently their original attitude was essentially horizontal. Further evidence for this is provided by banded chalcedony amygdales and partly filled vesicles with the "water level" parallel to the flow layering of the basalt (Plate 6E). The present high southerly dips of the Suurberg basalt in the Kirkwood-Paterson area are undoubtedly secondary. Formerly the relatively high dips of the Mesozoic rocks in this strip were ascribed to the presence of the "Zuurberg Fault". However, it is clear now that these dips were caused mainly by a subsiding basin or trough.

V SYNTHESIS AND CONCLUSIONS

The evidence presented in the descriptive portion of this work clearly shows that the Paleozoic and Mesozoic rocks of the Algoa Basin are separated by a major unconformity as a result of intervening oro= genesis and erosion (see p. 44); also that the Uitenhage rocks post= date the Suurberg volcanics with probably only a minor time lapse be= tween the two rock groups (p. 38). Thus, of the problems stated in the introductory paragraphs of Chapter I, only those concerning the nature, origin, and age of the Suurberg volcanism, and the initial development of the Algoa Basin deserve further consideration.

A. NATURE OF THE SUURBERG VOLCANISM

Evidence regarding the nature of the volcanism is provided by each of the rock units of the Suurberg Group.

1. The Slagboom Breccia

Provided a pyroclastic origin is accepted (cf. Chapter II, Sec= tion 1c), the following deductions can be made from the Slagboom breccia:

i) The Suurberg volcanism started with violent explosions, caused either by purely magmatic gases breaking through to the surface or, otherwise, by phreatic explosions. Such gas or steam explosions which do not expel essential ejecta are ordinarily of extreme violence and relatively low tem= perature (cf. A.G.I. Glossary, 1960, p. 219).

ii) The Slagboom breccia would indicate proximity of volcanic vents. Gorshkov and Dubik (1970) describe an exceptional case where volcanic blocks fell up to 10 km from an eruptive centre, but most blocks would fall much nearer to the vent.

iii) The continuous occurrence of breccia over a considerable distance indicates volcanic activity of a linear type, probably a fissure or a row of vents along a weak zone.

iv) The predominance of orthoquartzite in the breccia, to= gether with the paucity of tillite fragments, would suggest that the Witteberg quartzite formed the uppermost layer of the crust through which the gas and/or steam broke.

In connection with point (i) above the present author wants to revive an interesting possibility, first raised by A.W. Rogers, which might be of regional significance. Rogers (1906, p. 42) mentions a broad band of shattered rock, up to 1 km wide, which occurs in the Baviaanskloof, roughly 120 km northwest of Port Elizabeth. He was of the opinion that the scale of this brecciation could not be accoun= ted for by ordinary faulting alone and suggested volcanic explosions, possibly contemporaneous with the Suurberg volcanism, as a likely cause.

De Villiers (1941) also refers to the exceptional development of breccia in this region and writes: "The mass of shattered material is tremendous, thus on Elands Bosch there are hills over 1000 ft. high composed solely of breccia, while the zone parallel to the fault may be over half a mile wide. On examination the breccia is seen to con= sist of fragments of sandstones or quartzites set in a powder of similar material. These fragments are not always angular, frequently, e.g., on Couga, they are extremely well rounded, under which conditions it is often difficult to distinguish the breccia from a conglomerate. It is suggested that these breccias were formed as a result of the faulting of the rocks under insufficient cover, resulting in the large-scale shattering of the T.M.S. A second cause was, no doubt, due to the fact that faulting took place not in one plane only, but in several planes placed close together. Since movement probably did not take place along all the planes at the same time or with equal speed, rolling of the fragments would take place, with what may be termed a 'fault-conglomerate' as the final product" (op. cit., p. 157).

From the above description the lithological correspondence with the Slagboom Formation is evident. However, in the case of the latter a cataclastic origin is ruled out (p. 11). Hence, the question arises whether the Baviaanskloof breccias are indeed cataclastic in origin or whether they represent the products of volcanic gas or phreatic ex= plosions along a weak zone. In the author's opinion the possible genetic connection between the Baviaanskloof breccias and those of the Slagboom Formation warrants further investigation.

In many instances farther afield volcanism has started with vent-opening breccia like that of the fissure eruption of Threngslaborgir, Iceland, which is described in some detail by Rittmann (1962, p. 88-92). In the past some occurrences of vent-opening breccia have probably been overlooked because volcanism ceased before the extrusion of lavas or tephra could take place. In other instances vent-opening breccia has not yet been exposed by erosion owing to the fact that it occurs at the base of a volcanic sequence, and mostly only close to eruptive vents.

2. The Tuffs

The Suurberg tuffs contribute the following evidence regarding the nature of the volcanism:

i) The common occurrence of pisolitic tuff proves that the ejection took place subaerially (Moore and Peck, 1962, p. 191).

ii) The fact that these subaerial deposits have been preserved from destruction in an area of considerable relief indicates a rapid succession of depositional events.

iii) The pisolitic tuffs also indicate proximity of their source. Moore and Peck (1962, p. 191) conclude that accretionary lapilli would probably occur within 10 miles (16 km) from the associated eruptive vent.

iv) No important grain size variations amongst the tuffs are apparent when followed along their strike. This would suggest linear eruption. Furthermore, the occurrence of closely spaced eruptive centres on Beyers Vley Outspan (p. 17) in=dicates that ejective activity was probably localized at several points along a fissure or weak zone.

v) The many fragmented crystals and the generally fine-grained nature of the tuffs (cf. Table 3), together with the scarcity of pumice indicate a highly explosive eruption. This implies a gas-rich, acidic magma where, on rising to the surface, a gas phase separated out with a resultant sharp increase in viscosity. As the external pressure was further reduced expansion of the gas bubbles was inhibited by the high vis=cosity so that, eventually, their walls were violently rup=tured (Rittmann, 1962, p. 226).

The tuff eruptions were broadly of two types: ash-fall tuffs — indicated, amongst others, by the presence of accretionary lapilli (Moore and Peck, 1962); ash-flow tuffs — indicated by welding, axio=litic devitrification (Ross and Smith, 1961, p. 37), and gas tubes in certain tuffs (cf. p. 18). The indications for pyroclastic flow are usually found in tuffs from the Mimosa Formation and the upper portions of the Coerney Formation. Apparently most, if not all, of the older tuffs are the result of ash-falls. A distinction between the two types is not always possible because ash flows need not be welded.

Most workers on pyroclastic flows (ignimbrites) believe them to have been formed from a hot, rapidly expanding, turbulent, highly mobile, magmatic gas "cloud" (nuée ardente) which carries with it intratelluric crystals and liquid droplets of the exploding magma, as well as rock fragments torn from the walls of the vent (Cook, 1966, p. viii). The principal opposing hypothesis for a nuée ardente or ash-flow origin, advocated mainly by Russian authors, usually invokes highly gaseous lava flows. There can be little doubt that the Suurberg ignimbrites do not belong to this category because no "frothy" rocks were seen among them and their clastic nature is recognised at once in thin sec= tion.

Some of the main problems which usually surround ash flows are those concerning mode of transport and sufficient heat to cause welding of glass shards. Mobility of the ash flows is probably best explained by fluidization of the constituent particles (Reynolds, 1954, amongst others). Tuff veins, such as those found on Beyers Vley Outspan (see p. 17), are characteristically associated with fluidization vents (Reynolds, 1969, p. 111). The basaltic hornblende found in the tuff dykes of the above mentioned locality suggests that the temperature locally approached 800^{o}C (cf. p. 37), which is probably more than sufficient for welding of glass shards (Ross and Smith, 1961, p. 42).

3. The Basaltic Rocks

The basaltic rocks provide the following evidence regarding the nature of the volcanism:

i) The dolerite dykes of the Suurberg region probably represent feeder channels for the basalt and suggest that the latter extruded from local fissures.

ii) The absence or scarcity of pillow lavas, pipe amygdales, and megascopic linear features indicates quiet, subaerial flows.

iii) The distribution of the basalt on Mimosa (see p. 16) clearly shows that the flows were directed by the then existing to= pography. It follows that the regional elongate outcrop pattern of the Suurberg basalt, parallel to the strike of the underlying structure, relates to the same cause and that the distribution of basalt below the Uitenhage deposits would be confined to old valley floors.

iv) The high proportion of (glassy) groundmass (cf. Table 4) proves that the basalt cooled rapidly. This feature, to= gether with rather thin flow units and proximity of the ex= trusion vents, points to a relatively small scale extrusion of basalt. Further, when the paleotopography is considered, as well as the absence of large scale erosion of the top of the basalt (p. 21), it may be inferred that this latter unit probably did not extend much beyond its present limits.

4. Combined Evidence

When the different lines of reasoning are combined it becomes clear that the Suurberg extrusive activity was restricted to a small part of the denuded Cape Fold Belt. The eruptions took place sub=aerially from a number of vents along a weak zone or fissure in the following sequence:

i) Highly explosive (Plinian) opening phase producing vent-opening breccia followed by ash falls.

ii) Extrusion of ash flows.

iii) Quiet effusion of basalt, temporarily interrupted by minor ash flows.

There is an overall decrease of explosivity which reflects the extrusion of highly viscous acidic magma coming from progressively deeper levels and having correspondingly lower gas content, followed by much less viscous basic magma.

B. POSSIBLE ORIGIN OF THE SUURBERG MAGMAS

Perhaps one of the most important facts which has emerged from the present study is the occurrence in the Suurberg area of two dis=tinct igneous rock classes — rhyolitic and basaltic — while there appears to be a complete absence of intermediate types.

The Suurberg basalt, it was shown (p. 42), is a typical tholeiite. In the light of recent experimental work this would imply that it was derived from the upper mantle by partial melting and crystal fractiona=tion (cf. Green, 1970). Although the composition of the parental upper mantle can only be inferred the pyrolite (pyroxene + olivine rock) source of Green and Ringwood (1967) currently appears to provide an acceptable working model.

The origin of the rhyolitic magma appears to be less straight=forward. Current hypotheses would suggest mainly two sources: i) Differentiation of a basaltic parent magma generated in the mantle; and ii) melting or partial melting of crustal rocks (anatexis). Point i) might also involve subsequent assimilation of sialic materials during ascent of the magma. There are many other smaller variations like that of Yoder (1969, p. 85), for instance, who shows that small amounts of rhyolitic magma might be produced by partial fractional remelting of quartz-normative, hydrous andesite.

However, except for point ii) above these various modes of origin are all discounted by the absence of intermediate rock types, as well as the apparently high "rhyolite": basalt ratio in the Suurberg area.

Assuming only minor erosion of the basalt (see p. 21) and absence of more basalt in depth (see later), this ratio works out at approximate= ly 1,15:1 when only the stratigraphic columns of Figs. 2 and 3 are taken into account. In reality this proportion is probably even higher be= cause the larger lateral extent of the tuffs is not properly allowed for in the stratigraphic columns.

At the same time the relatively high values of silica and alumina of the acidic Suurberg rocks (see Table 10) would indicate derivation from the sialic crust. This is further supported by experimental studies on the crystallization of feldspar which suggest that silica-rich, two-feldspar rhyolites are products of fusion of sialic material (Carmichael, 1963). Crystallization differentiation appears to be quite inadequate to account for silica-rich rhyolites (such as the Suurberg acidic rocks), because their silica content is significant= ly higher than that represented by the liquidus minimum in the silica-alkali feldspar system (op. cit., p. 110).

There appears to be a number of ways to obtain the heat required for the generation of anatectic magmas (cf. Middlemost, 1971, p. 116):

i) Frictional heating such as produced by shearing or by one crustal plate being thrust under another;

ii) the emplacement of high temperature magma into the crust;

iii) the lowering of crustal materials into a higher tempera= ture environment;

iv) the stress-release hypothesis which supposes that the materials within the earth have been locally remelted by a drop in melting temperature, below the surrounding tem= perature, upon release of pressure (Uffen and Jessop, 1963).

All of the above factors are usually associated with some or other stage of the orogenic cycle. Table 12 (modified from Rittmann, 1962, p. 163) shows the general relationship between the different phases of orogenesis and the corresponding magmatism.

TABLE 12

PHASE OF OROGENIC CYCLE	CORRESPONDING MAGMATISM
Post-orogenesis (denudation)	Post-orogenic or final volcanicity (effusive basaltic or ignimbritic)
Orogenesis *sensu stricto* (uplift)	Orogenic to late orogenic or subse= quent volcanicity (explosive Pacific)
Tectogenesis (folding)	Tectogenetic or synorogenic plutonism (intrusive acid Pacific)
Geosynclinal	Geosynclinal or initial volcanicity (effusive basaltic-picritic)

The volcanism of the Suurberg region is obviously connected with the last stages of the orogenic cycle, but it should also be borne in mind that the time lapse between the generation of magmas and their actual eruption can be considerable (cf. Rittmann, 1962, p. 216). It is likely that anatexis of crustal rocks in the Cape Fold Belt took place during tectogenesis resulting, locally, in diapiric intrusions of acid magmas. Should crystallization of these magmas have taken place, remelting could have been effected by subsequent release of stress. At the same time the formation of tectonic fissures would enable the rhyolitic magma to be erupted at the surface.

Uffen and Jessop (1963) have shown that, in association with earthquakes, granitic magmas may be generated between depths of approxi= mately 15 and 50 km. For basaltic magmas the corresponding depths are between 50 and 100 km. The tectonic earthquakes provide evidence of the release of stresses within the earth and are often particularly marked in the period immediately following orogenesis.

It is further significant that the Suurberg volcanics occur more or less in the area where units of the Cape Supergroup appear to reach maximum thickness (e.g., Theron, 1970, Fig. 4). Immediately south and west of the Kirkwood Panhandle Johnson (1966) has calculated a possible total thickness in the order of 10 000 metres for the units between the base of the Table Mountain Group and the top of the Main Witteberg. In the Willowmore district, farther towards the west, Theron (1962) has used a corresponding figure of 3 300 metres in con= structions to calculate the depth of folding. Provided then that the intensity of folding is of the same order in the two areas, which appears to be the case, it is clear that the depth of folding in the vicinity of the present Algoa Basin should be considerably more than the mini= mum of 16 km calculated by Theron for the Cape rocks of the Willowmore area (op. cit., p. 402). The folding certainly also affected a further unknown thickness of late Precambrian clastic rocks, mainly quartzite, arkose, phyllite, and limestone, which underlie the Table Mountain Group (Amm. 1935).

Disregarding, for the moment, possible later differential uplift, a further indication that maximum depression of the sedimentary rocks took place in the Algoa area is afforded by the remnants of infolded Ecca rocks (p. 9). Elsewhere in the Cape Fold Belt the Ecca has long since been stripped off, appearing only on the north side of the main fold ranges where, due to rapid wedging northwards, the Cape rocks are also considerably thinner.

Anatectic magmas, derived from mountain roots, are usually rich in volatiles and this is expressed by their highly explosive extru= sion. Basaltic magmas, although higher in total energy content, do not possess the same eruptive power as gaseous acidic magmas (Rittmann, 1962, p. 218). For this reason the former is dependent on tensional stresses to create passages to the surface, whereas acidic magmas, under favourable circumstances, might rupture the roof due to internal pressure build-up. It would appear as if a weak zone, which had de= veloped through faulting in the Algoa area, was opened sufficiently by the explosive action of a gas-rich acidic magma, to enable basal= tic magma to gain access to the surface. This local tension (faul= ting), combined with the vent drilling capacity of a rhyolitic magma, would firstly explain the presence of basalt locally within the Cape Fold Ranges where, otherwise, conditions appear to have been unfavour= able for basaltic intrusions (cf. the distribution of Karoo dolerites whose southerly limit seems to be controlled by the Fold Belt). Second= ly, this would also explain why the basalt overlies the rhyolitic tuffs. Generally, where contemporaneous eruptions of basic and acidic magmas occur, the former are slightly earlier, possibly on account of lower viscosity and greater mobility (King, 1965, p. 445).

The absence in the Suurberg region of intermediate rock types, more specifically andesites, deserves some comment since they are so commonly associated with active continental margins and orogenic re= gions elsewhere, e.g. the Andes of South America. Several hypotheses regarding the origin of andesitic rocks have been proposed (cf. Middle= most, 1971, p. 122). If andesites formed as a result of thorough mingling of acidic and basic magmas in depth (Larsen et al., 1938, p. 429), circumstances appear to have been favourable for their formation in the Suurberg area. On the other hand, Yoder (1971) provides ex= perimental evidence to show that rhyolitic and basaltic magmas could possibly exist in physical contact without any appreciable mixing taking place.

Modern opinion seems to favour the idea that andesitic magmas are generated where oceanic crustal materials are thrust into the mantle beneath continental margins, this process being intimately connected with the orogenic cycle (cf. McBirney, 1970, p. 346). At first, this makes the absence of andesites in the Suurberg region even more sur= prising. However, a similar situation is found in the Alpine orogenic belt of Southern Europe. Superficially it would appear that where two continental plates are involved in the mountain building (the Alps) andesites are absent, while they occur abundantly in areas where a continental plate collides with an oceanic crustal plate (the Andes). In the case of the Suurberg region this explanation seems to have some merit in the light of strong evidence that the different parts of Gondwanaland must have been close together during the time that the Cape Folding took place (Carboniferous to Triassic).

Perhaps another approach for a solution is offered by Yoder's (1969) suggestion that andesites are the products of partial melting, under hydrous conditions, of parental material which, under anhydrous conditions, would have yielded tholeiitic basalt. This concept would explain the usual paucity of basalt associated with calc-alkaline andesites in continental margins (op. cit., p. 85), and conceivably also the reverse situation.

To summarize, it is clear that the Suurberg rhyolitic and basal= tic magmas formed independently from different sources and possibly also at different stages of the orogenic cycle. Their contempora= neous eruption was due to tectonic causes which created passageways towards the surface; the volatile-rich anatectic magma possibly played an auxiliary role by clearing the vents sufficiently to allow ascent of basalt from below.

C. AGE AND GENETIC RELATIONSHIPS OF THE SUURBERG VOLCANISM

To date only one radiometric dating of a Suurberg rock is availa= ble (see p. 3). The figure obtained, 80 to 90 m.y., appears to be at variance with available paleontological and other field data.

The direct field evidence shows that the Suurberg volcanic rocks are considerably younger than the Lower Ecca (Permian) from which they are separated by a major unconformity, and slightly older than the Enon Formation of the Uitenhage Group. It is also clear that the volcanism postdates the major pulse of the Cape Folding which occurred during the Early to Mid Triassic.

The fossil material found in the Suurberg rocks (p. 15) has so far been unsuitable for age determination. Hence, for an upper age limit the overlying Enon is of utmost importance. The age of the Enon rocks has long been taken to be Late Jurassic or Early Cretaceous (Du Toit, 1954, p. 382), based on the fossil content of the overlying Wood Beds and Sundays River Formation. The paleontological ages of the Uitenhage rocks have recently been subjected to more detailed investigation on be= half of Soekor. The results obtained by different laboratories using ostracods, foraminifera, ammonites, and plant microfossils do not always agree mutually but, on the whole, tend to confirm the earlier accepted age (Late Jurassic - Early Cretaceous) for the lowermost Uitenhage rocks (Dr. P.J. van Zijl, 1971, personal communication).

Thus, at this stage, it would appear reasonable to infer a (Mid to Late) Jurassic age for the rocks of the Suurberg Group.

Cox (1970, p. 219-220) points out that, based on radiometric dating, igneous activity seems to have been in progress in one part or another of Southern Africa throughout the entire time-range from 200 to 100 million years (Late Triassic to Mid Cretaceous). However, for the sake of convenience, he restricts the term "Karroo" to those lavas of Southern Africa which are of Jurassic age. According to this defini= tion the Suurberg volcanic rocks should also be included with the Karoo lavas.

The petrographic and chemical affinity of the Suurberg basalts with those of the Southern Basalt Province of Cox et al. (1966) has already been pointed out in Chapter III. On the other hand, the Suurberg rhyolitic rocks differ considerably from their nearest geographical counterparts in the Lebombo area (cf. Table 10). Also, in the latter region all stages of intermediate rocks between basaltic and rhyolitic are found. This fact, together with work on strontium isotope ratios (Manton, 1968), would indicate that the Lebombo acid and intermediate rocks were derived from a magma which originated in the upper mantle and became modified through fractional crystallization and assimilation of crustal materials (Stratten, 1965). The heat for this latter process was probably supplied by a large linear wedge of basic magma, indicated by a high positive gravity anomaly (Smit et al., 1962). In the Suurberg area, on the other hand, the existence of a negative gravity anomaly discounts the presence, in depth, of larger basic bodies. A further difference is that the eruption of the main body of rhyolitic rocks in the Lebombo area took place mainly after extrusion of basic lavas, while the reverse order ob= tains in the Suurberg region.

In contrast to the obvious differences in chemical composition and tectonic setting between the Lebombo and Suurberg acidic rocks, the latter have much closer correspondence with the alkali-rhyolitic rocks of the Andes which are also regarded as products of anatexis (Zeil and Pichler, 1967, p. 77). There is one important difference though: The normative feldspars and quartz of the Andean rocks occur in more or less eutectic proportions, while those of the Suurberg tuffs show a large excess of quartz (see Table 10). It is therefore suggested that the magma which gave rise to the acidic Suurberg rocks had been derived through anatexis of quartzose clastic rocks rather than remelting of granitic parent mate= rial. This deduction seems justified in the light of the extraordinarily thick sediment accumulation and the large-scale folding which took place here (see p. 55). In these respects the Cape Fold Belt surpasses the Andes where, in contrast, igneous activity assumed a much more promi= nent role (cf. Zeil and Pichler, 1967, p. 79).

The apparent genetic relationship between the Suurberg volcanism and the Cape orogenic cycle has already been pointed out in the previous section. The pre-Uitenhage faulting (p. 61) and Suurberg volcanicity are clearly connected with the post-orogenic tensile stresses which must have prevailed at least during most of the Jurassic (cf. De Villiers, 1944, p. 206). Walker and Poldervaart (1949, p. 690) suggest that the

Suurberg basalt was a direct result of the syntaxis of the great Natal-Lebombo Monocline and the Samfrau Orogenic Belt (of which the Cape Fold Belt supposedly forms part). Therefore they consider that the east-west faults in the Cape Fold Belt probably initiated in Trias-Lias times. Local post-Enon characteristics of the faults could well be ascribed to later rejuvenation. On the other hand, Söhnge (1934) convincingly demonstrated the post-Enon age of at least 1 - 2 km of movement along the Worcester Fault farther towards the west. Assuming contem=poraneity of the Enon deposits, this would suggest that the maximum in=tensity of tensile stresses during the Mesozoic probably shifted pro=gressively from east to west along the southern Cape Fold Belt. The further implication is clear, namely, that the supposed rifting of Gondwanaland parallel to the southerly South African coast line during Mesozoic times probably progressed from east to west.

D. EARLY BASIN HISTORY

During the Suurberg volcanicity probably no true tectonic basin, i.e. one directly controlled by crustal movements, existed in the nor=thern Algoa region. The volcanic products simply filled longitudinal valleys which had been carved out in the fold ranges under control of the underlying lithology and structure, including pre-Uitenhage faulting. This erosion was mainly the result of the last major orogenic pulse of the Early to Mid Triassic (De Villiers, 1944, p. 200).

Except for certain conglomerates in the Slagboom Formation, prac=tically no other externally derived rudites were found interbedded with the Suurberg lavas and tuffs. This is not easily understood, consider=ing the topography at the time and the relatively thick succession of Enon conglomerates following immediately upon the volcanic rocks. It would, at any rate, corroborate other evidence that the Suurberg volca=nism was a rapid, short-lived event (cf. p. 51). Further, it would also indicate that some rather drastic physiographic change followed the ex=trusion of the lavas.

The present author believes that a subsiding basin developed only at the close of the Suurberg volcanism. Two general points are rele=vant to such an event:

Firstly, local sinking of the crust often follows certain types of volcanism, thereby compensating for the subsurface mass deficiency caused by the extrusions. Ignimbrites, in particular, are commonly associated with caldera-like collapse features (McBirney, 1970, p. 348). The displacement of materials as a result of the Suurberg volcanism was relatively small, so that it probably achieved no more than "triggering off" further crustal subsidence. It is notable that the Algoa basining took place in an area where formerly, during the deposition of the

Paleozoic rocks, subsidence appeared to have been maximal, thus in= dicating an inherently weak part of the crust.

Secondly, there is evidence that the Suurberg volcanic units, which follow concordantly one upon the other, came to rest essential= ly in a horizontal attitude (see p. 48). Their present dips must have been caused by subsequent sinking of the basin. A sympathetic relationship exists between the magnitude of these dips and the de= velopment of Enon conglomerate along the inner margin of the northern Algoa Basin (see p. 48), which could hardly have been accidental. It is suggested that the higher dips found along certain parts of the basin margin reflect stronger initial subsidence rather than increasing rates of sinking at a later stage. On these comparatively steeper slopes only coarser rudites, derived from higher terrain, accumulated (owing to higher inertia) while finer clastics were washed down into the deeper parts of the basin. This extraordinary situation might also have been responsible for the unusual imbrication which occurs locally in the basal Enon conglomerates (see p. 20).

Along the periphery of the Northern Algoa Basin this early sinking must have had maximum effect in the Kirkwood-Paterson area (see Table 11). It is suggested therefore that the initial subsidence of the northern part of the basin was accomplished by downwarping towards the south. A downwarp with its east-west axis situated far enough north would have caused a transgression of the Enon over the older Suurberg rocks. Such actually appears to be the case in the field. A transgressive Enon sedimentation, coupled with restricted basalt distribution (p. 52) also seems to be the only way of explaining the absence of basalt clasts in the lowermost Enon conglomerate (see pp. 22, 38).

Further subsidence of the basin appears to have been rather ir= regular, perhaps partly due to contemporaneous faulting. The central portions of the basin were mainly affected so that temporary lowering of base level laid bare earlier deposits along the edge of the deposi= tory. In this way local erosion of these earlier units could have supplied the basalt clasts found higher up in the Enon (cf. p. 20). However, the total effect of the Uitenhage deposition was transgressive. This is clearly demonstrated by the eventual marine incursions towards the top of this succession.

VI SYNOPSIS OF GEOLOGICAL HISTORY

The deductions of the foregoing chapter will now be summarized in chronological order and related to the geological events which af= fected the area since Paleozoic times. Data regarding the ages of deformation and the regional geological developments are mainly from De Villiers (1944) and Haughton (1969, p. 487-497).

Early Paleozoic to Devonian: Erosion of a slowly rising conti= nental interior supplied detritus which was deposited towards the south on an east-west trending marginal shelf to give rise to the rocks of the Cape Supergroup. The rate of subsidence of this depository in= creased eastwards.

Carboniferous: During the Early Carboniferous, deposition on the Cape Shelf ended when the first pulse of (mild) tectonic deforma= tion affected the Cape rocks. At the same time a latitudinal trough was initiated along the southern fringe of the Karoo (cf. geosynclinal phase of Table 12). The earlier deposits of this trough came from the north and south and were mainly glacially transported.

Permian: For most of this period Ecca sediments accumulated in the geosynclinal trough. Detritus was derived from the east, south, and west. During the Late Permian a second deformational pulse affected the deposits within the trough (cf. Tectogenesis of Table 12). At this stage anatectic magmas could have started to develop where crustal materials reached down into a high temperature environment, especially in the east where maximum subsidence had oc= curred. The main locus of sedimentation within the greater Karoo Basin gradually shifted northwards while continental sediments of the Beaufort Group were being deposited.

Triassic: The third, and probably most important, deformational pulse to affect the Paleozoic rocks of the southern Cape occurred during the Early to Mid Triassic epochs and culminated in the development of the Cape Fold Belt by large scale folding, thrusting, and uplift (cf. Orogenesis, sensu stricto, of Table 12). The denudation which followed further shaped the southern Cape mountain ranges, including those of the Suurberg and Winterhoek. The products of this erosion accumulated farther north in the central Karoo Basin. Towards the close of the Triassic tensile forces replaced the regional compres= sive stress which had formerly played such a dominant role, and the first manifestations of volcanism within the Karoo Basin became evi= dent.

Jurassic: The main eruption of widespread tholeiitic basalts and dolerites which terminated the deposition of Karoo rocks occurred during the Early Jurassic. Tensional stresses also prevailed in the Klein Winterhoek-Suurberg region of the Cape Fold Belt, probably in response to the first stages of separation of Gondwanaland along the present southern Cape coast. Local relief through normal faulting en= abled the earlier formed volatile-rich anatectic magma to rise and break through the formations of the Cape Supergroup. This took place along a weak (faulted) zone in the form of violent gas explosions. The opening phase was followed first by ash-fall tuffs, produced by violent disruption of the highly viscous magma, and then by more re= stricted ash flows. The rhyolitic volcanism was succeeded by tho= leiitic basalt flows which exploited the newly created passageways to ascend through the thick, competent crust. The lava flows were re= stricted mainly to intermontane valleys. The crust started to sub= side soon after the Suurberg volcanism ended, probably through down= warping, so that the Algoa Basin took form as a separate tectonic entity. These crustal movements during the Late Jurassic also ini= tiated the deposition and transgression of the torrential Enon de= posits over the volcanic rocks.

Cretaceous to Recent: With further subsidence and sedimentation during the Early Cretaceous the Algoa Basin developed to its full ex= tent. The absence of major post-Uitenhage folding or faulting in the Northern Algoa Basin suggests that subsequent compressional or ten= sional stresses never reached critical proportions in this region. The later history of the Algoa area up to the present can be deduced from spasmodically preserved deposits that reflect epeirogenic move= ments at different times, coupled with eustatic rise and fall of sea level. The nett result was considerable elevation of the land during the Cainozoic, followed by prolonged erosion which eventually also exposed the volcanic rocks of the Northern Algoa Basin.

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REFERENCES

AMERICAN GEOLOGICAL INSTITUTE, 1960. Glossary of geology and re= lated sciences (2nd ed.). Washington, D.C.

AMM, F.L., 1934. The Pre-Cape rocks of the Uitenhage district. Trans. geol. Soc. S. Afr., 37, 69-86.

ANHAEUSSER, C.R., MASON, R., VILJOEN, M.J., and VILJOEN, R.P., 1968. A reappraisal of some aspects of Precambrian shield geolo= gy. Econ. Geol. Res. Unit, Univ. Witwatersrand, Inf. Circ. 49.

BILLINGS, M.P., 1954. Structural Geology (2nd ed.). Prentice-Hall, Englewood Cliffs, N.Y.

BOTHA, B.J.V. and THERON, J.C., 1967. New evidence for the early commencement of Stormberg volcanism. Tydskrif vir Natuurweten= skappe, 7, 469-473.

CARMICHAEL, I.S.E., 1963. The crystallization of Feldspar in volca= nic acid liquids. Quart. J. geol. Soc. Lond., 119, 95-131.

CHAYES, F., and ZIES, E.G., 1961. Staining of alkali feldspars from volcanic rocks. Carnegie Inst. Wash. Year Book, 60, 172-173.

COOK, E.F., 1966. Tufflavas and ignimbrites. Elsevier, New York.

COX, K.G., 1970. Tectonics and volcanism of the Karoo Period and their bearing on the postulated fragmentation of Gondwanaland. In: African magmatism and tectonics, 211-235. Oliver and Boyd, Edinburgh.

―――――, and HORNUNG, G., 1966. The petrology of the Karroo basalts of Basutoland. Amer. Min., 51, 1414-1432.

―――――, MACDONALD, R., and HORNUNG, G., 1967. Geochemical and petrographic provinces in the Karroo basalts of Southern Africa. Amer. Min., 52, 1451-1474.

DEER, W.A., HOWIE, R.A., and ZUSSMAN, J., 1966. An introduction to the rock-forming minerals. Longmans, London.

DE VILLIERS, J., 1944. A review of the Cape Orogeny. Ann. Univ. Stellenbosch, 22 (A, 10), 183-208.

DE VILLIERS, J.E., and WARDAUGH, T.G., 1962. A sedimentary petro= logical study of some sandstones, conglomerates and tillites of the Cape and Karroo Systems. Trans. geol. Soc. S. Afr., 65, 101-108.

DRYDEN, L., and DRYDEN, C., 1946. Comparative rates of weathering of some common heavy minerals. J. sediment. Pet., 16, 91-96.

DU TOIT, A.L., 1930. The volcanic belt of the Lebombo; a region of tension. Trans. roy. Soc. S. Afr., 18, 189-218.

―――――, 1954. The geology of South Africa (3rd ed.). Oliver and Boyd, Edinburgh.

ENGELBRECHT, L.N.J., COERTZE, F.J., and SNYMAN, A.A., 1962. Die geologie van die gebied tussen Port Elizabeth en Alexandria, Kaapprovinsie. Expl. sheet 3326C (Alexandria), geol. Surv. S. Afr.

FOLK, R.L., 1968. Petrology of sedimentary rocks. Hemphill's, Austin, Texas.

FRANKEL, J.J., 1960. The geology along the Umfolosi River, south of Mtubatuba, Zululand. Trans. geol. Soc. S. Afr., 63, 231-252.

GEORGE, W.O., 1924. The relation of the physical properties of natural glasses to their chemical composition. J. Geol., 32, 353-372.

GORSHKOV, G.S., and DUBIK, Y.M., 1970. Gigantic directed blast at Shiveluch Volcano (Kamchatka). Bull. volcan., 34, 261-288.

GOTTINI, V., 1968. The TiO_2 frequency in volcanic rocks. Geol. Rdsch., 57, 930-935.

——————, 1969. Serial character of the volcanic rocks of Pantelleria. Bull. volcan., 33, 818-827.

GRIM, R.E., 1968. Clay mineralogy (2nd ed.). McGraw-Hill, New York.

GREEN, D.H., 1970. A review of experimental evidence on the origin of basaltic and nephelinitic magmas. Phys. Earth Planet. In= teriors, 3, 221-235.

——————, and RINGWOOD, A.E., 1967. The genesis of basaltic magmas. Contr. Mineral. Pet., 15, 103-190.

GUEST, J.E. and JONES, G.P., 1970. Origin of ash deposits in the Santiago area, Central Chili. Geol. Mag., 107, 369-381.

HATCH, F.H. and CORSTORPHINE, G.S., 1909. The geology of South Africa (2nd. ed.). MacMillan and Co., London.

HAUGHTON, S.H., 1928. The geology of the country between Grahams= town and Port Elizabeth. Expl. Cape sheet 9 (Port Elizabeth), geol. Surv. S. Afr.

——————, 1935. The geology of the country east of Steytler= ville. Expl. sheet 150 (Sundays River), geol. Surv. S. Afr.

——————, 1969. Geological history of Southern Africa. Geol. Soc. S. Afr., Johannesburg.

——————, and ROGERS, A.W., 1924. The volcanic rocks south of Zuurberg. Trans. roy. Soc. S. Afr., 11, 235-248.

HUGHES, C.J., 1960. The southern mountains igneous complex, Isle of Rhum. Quart. J. geol. Soc. Lond., 116, 111-138.

JOHNSON, M.R., 1966. The stratigraphy of the Cape and Karroo systems in the Eastern Cape Province. Unpubl. M.Sc. thesis, Rhodes Univ., Grahamstown.

KELLER, J., 1969. Origin of rhyolites by anatectic melting of gra= nitic crustal rocks: Bull. volcan., 33, 942-959.

KING, B.C., 1965. The nature of basic igneous rocks and their relations with associated acid rocks. Sci. Progr., 53, 437-446.

KRUMBEIN, W.C., 1940. Flood gravels of San Gabriel Canyon, Califor= nia. Bull. geol. Soc. Amer., 51, 639-676.

——————, 1942. Flood deposits of Arroyo Seco, Los Angeles county, California. Bull. geol. Soc. Amer., 53, 1355-1402.

——————, and SLOSS, L.L., 1963. Stratigraphy and sedimen= tation (2nd ed.). W.H. Freeman & Co., San Francisco.

LARSEN, E.S., IRVING, J., GONYER, F.A., 1938. Petrologic results of a study of the minerals from the Tertiary volcanic rocks of the San Juan region, Colorado. Am. Miner., 23, 417-429.

LOOCK, J.C., 1967. The stratigraphy of the Witteberg-Dwyka contact beds. Unpubl. M.Sc. thesis, Univ. Stellenbosch, Stellenbosch.

MACKENZIE, W.S., and SMITH, J.V., 1956. The alkali feldspars (Part III). Amer. Min., 41, 405-427.

MANSON, V., 1967. Geochemistry of basaltic rocks: major elements. In: Basalts, Volume 1, 215-269. Interscience Publishers, New York.

MANTON, W.I., 1968. The origin of associated basic and acid rocks in the Lebombo-Nuanetsi igneous province, Southern Africa, as im= plied by strontium isotopes. J. Pet., 9, 23-39.

McBIRNEY, A.R., 1970. Some current aspects of volcanology. Earth-Sci. Rev., 6, 337-352.

MEYER, W., 1965. The geology of a portion of South-western Albany. Unpubl. M.Sc. thesis, Rhodes Univ., Grahamstown.

MIDDLEMOST, E.A.K., 1971. Classification and origin of the igneous rocks. Lithos, 4, 105-130.

MOORE, J.G., and PECK, D.L., 1962. Accretionary lapilli in vol= canic rocks of the Western Continental United States: J. Geol., 70, 182-193.

MOUNTAIN, E.D., 1946. The geology of an area east of Grahamstown. Expl. sheet 136, geol. Surv. S. Afr.

MURAI, I., 1963. Pyroclastic flow deposits on various volcanoes in Japan. Bull. volcan., 26, 337-351.

NOCKOLDS, S.R., 1954. Average chemical compositions of some igneous rocks. Geol. Soc. Amer. Bull., 65, 1007-1032.

PANDE, I.C., and GUPTA, V.J., 1965. Occurrence of "explosion brecciae" at Drang, Mandi district, Himachal Pradesh, India. Current Science, 34, 315-316.

PETERSON, D.W., and ROBERTS, R.J., 1963. Relation between the crystal content and the chemical composition of welded tuffs. Bull. volcan., 26, 113-123.

PETTIJOHN, F.J., 1957. Sedimentary rocks. (2nd ed.). Harper & Brothers, New York.

PINCHIN, R., 1875. A short description of the geology of part of the Eastern Province of the Colony of the Cape of Good Hope. Quart. J. geol. Soc. Lond., 31, 106-108.

PITTMAN, E.D., 1970. Plagioclase feldspar as an indicator of pro= venance in sedimentary rocks. J. sediment. Pet., 40, 591-598.

POTTER, P.E., and PETTIJOHN, F.J., 1963. Paleocurrents and basin analysis. Springer-Verlag, Berlin.

PRICE, N.J., 1966. Fault and joint development in brittle and semi-brittle rock. Pergamon Press, Oxford.

REYNOLDS, D.L., 1954. Fluidization as a geological process, and its bearing on the problem of intrusive granites. Amer. J. Sci., 252, 577-613.

——————, 1969. Fluidization as a volcanological agent. Proc. geol. Soc. Lond., no. 1655, 110-113.

RIGASSI, D., 1968. Preliminary report on the geology and oil pros= pects of the Sunday's River Basin. Unpubl. Rep. Petroconsultants S.A., Geneva.

RITTMANN, A., 1954. Remarks on the eruptive mechanism of the Ter= tiary volcanoes of Egypt. Bull. volcan., 15, 109-117.

——————, 1962. Volcanoes and their activity. John Wiley & Sons, New York.

——————, 1967. Die Bimodalität des Vulkanismus und die Her= kunft der Magmen. Geol. Rdsch., 57, 277-295.

ROGERS, A.W., 1905. The volcanic fissure under Zuurberg. Trans. phil. Soc. S. Afr., 16, 190-197.

——————, 1906. Geological Survey of parts of the Divisions of Uitenhage and Alexandria. Annu. Rep. geol. Comm. C. G. H. for 1905, 12-42.

ROSS, C.S., and SMITH, R.L., 1961. Ash-flow tuffs: Their origin, geologic relations and identification. U. S. geol. Surv. Prof. Pap. no. 366.

RUDDOCK, A., 1968. Cainozoic sea-levels and diastrophism in a region bordering Algoa Bay. Trans. geol. Soc. S. Afr., 71, 209-233.

SMIT, P.J., HALES, A.L., and GOUGH, D.I., 1962. Gravity survey of the Republic of South Africa. Handb. geol. Surv. S. Afr., no. 3.

SÖHNGE, P.G., 1934. The Worcester Fault. Trans. geol. Soc. S. Afr., 37, 253-277.

STRATTEN, T., 1965. Pyroclastic and associated igneous rocks of the Lebombo Mountain Range south of the Great Usutu River, Zululand. Unpubl. M. Sc. thesis, Univ. Potchefstroom, Potchefstroom.

STRECKEISEN, A., 1966. Die Klassifikation der Eruptivgesteine. Geol. Rdsch., 55, 478-491.

THERON, J.N., 1962. An analysis of the Cape Folding in the district of Willowmore, C. P. Ann. Univ. Stellenbosch, 37, (A, 5), 347-419.

——————, 1970. A stratigraphical study of the Bokkeveld Group. Proc. 2nd I.U.G.S. Gondwana Symp., South Africa, 197-204.

THORNTON, C.P., and TUTTLE, O.F., 1960. Chemistry of igneous rocks: Differentiation index. Amer. J. Sci., 258, 664-684.

TOBI, A.C., 1956. A chart for measurement of optic axial angles. Amer. Min., 41, 516-519.

TRUSWELL, J.F., 1967. A critical review of stratigraphic terminology as applied in South Africa. Trans. geol. Soc. S. Afr., 70, 81-116.

——————, 1970. An introduction to the historical geology of South Africa. Purnell & Sons, Cape Town.

——————, and RYAN, P.J., 1969. A flysch facies in the lower Ecca Group of the southern Karroo and a portion of the Transkei. Trans. geol. Soc. S. Afr., 72, 151-158.

UFFEN, R.J., and JESSOP, A.M., 1963. The stress release hypothesis of magma formation. Bull. volcan., 26, 57-66.

VAN BEMMELEN, R.W., 1954. Mountain building. Martinus Nijhoff, The Hague.

————————, 1968. On the origin and evolution of the earth's crust and magmas. Geol. Rdsch., 57, 657-705.

VENTER, J.J., 1969. Stratigraphy and correlation of the Cape and Karroo Supergroups in the southern Cape Province. Unpubl. Rep. Soekor, Johannesburg.

WALKER, F., and POLDERVAART, A., 1949. Karroo dolerites of the Union of South Africa. Geol. Soc. Amer. Bull., 60, 591-706.

WILKINSON, J.F.G., 1967. The petrography of basaltic rocks. In: Basalts, volume 1, 163-214. Interscience Publishers, New York.

WILLIAMS, P.L., 1960. A stained slice method for rapid determination of phenocryst composition of volcanic rocks. Amer. J. Sci., 258, 148-152.

WINTER, H. de la R., and VENTER, J.J., 1970. Lithostratigraphic correlation of recent deep boreholes in the Karroo-Cape Sequence. Proc. 2nd I.U.G.S. Gondwana Symp., South Africa, 395-408.

WISE, W.S., and EUGSTER, H.P., 1964. Celadonite: synthesis, thermal stability and occurrence. Amer. Min., 49, 1031-1083.

YODER, H.S., 1969. Calcalkalic andesites: experimental data bearing on the origin of their assumed characteristics. In: Proc. Andesite Conf., Oregon Dept. Geol. & Min. Ind., Bull. 65, 77-89.

————————, 1971. Contemporaneous rhyolite and basalt. Car= negie Inst. Washington, Ann. Rept. Director Geophys. Lab. 1969-1970, 141-145.

————————, and TILLEY, C.E., 1962. Origin of basalt magmas. J. Petrology, 3, 342-532.

ZEIL, W., and PICHLER, H., 1967. Die känozoische Rhyolith-Formation im mittleren Abschnitt der Anden. Geol. Rdsch., 57, 48-81.

PLATE 1

A Quartzite breccia (Slagboom Formation) exposed in a gully on Drie Kuilen, Steytlerville district.

B Dinosaur bone fragment (?) occurring in a welded tuff on Gorie Laaghte, Uitenhage district.

C Non-welded tuff which occurs above the basalt on Rhinoster Hoek no. 2, Uitenhage district. The vertical tubes were probably formed by escaping gas.

D A portion of a thin sheet of partly welded tuff underlying the basalt. Looking east on Gorie Laaghte, Uitenhage dis= trict.

E Ash-fall tuff with black accretionary lapilli inclusions, some showing concentric "shells". Slag Boom, Kirkwood district.

F Bedded ash-fall tuff with a mottled appearance. Slag Boom, Kirkwood district.

G Welded tuff (light) which is intrusive into vesicular basalt (dark) of a volcanic "neck" on Beyers Vley Outspan, Steytler= ville district. The tuff dyke encloses xenoliths of the host rock. (See also Plates 2C, 6A & B).

PLATE 1

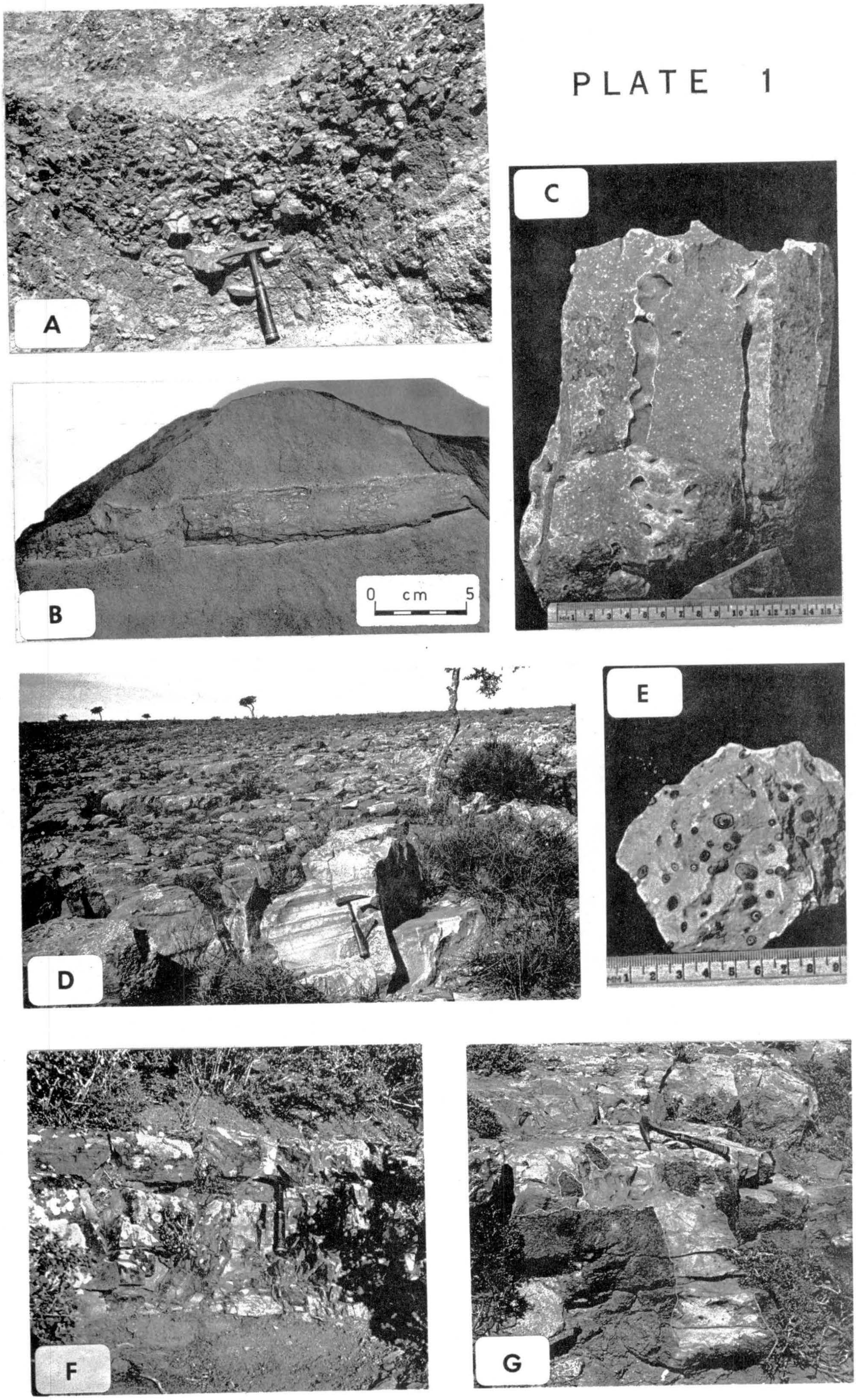


PLATE 2

A Contact between basalt (a) and basal Enon sandstone and shales (c). The intermediate sediments (b) are highly disturbed, probably due to faulting or gliding along the contact plane. Looking east on Beyers Vley Outspan, Steytlerville district.

B Thinly bedded basal Enon sandstone and shales on Beyers Vley Outspan. The folding of the beds is probably due to slumping. Looking east.

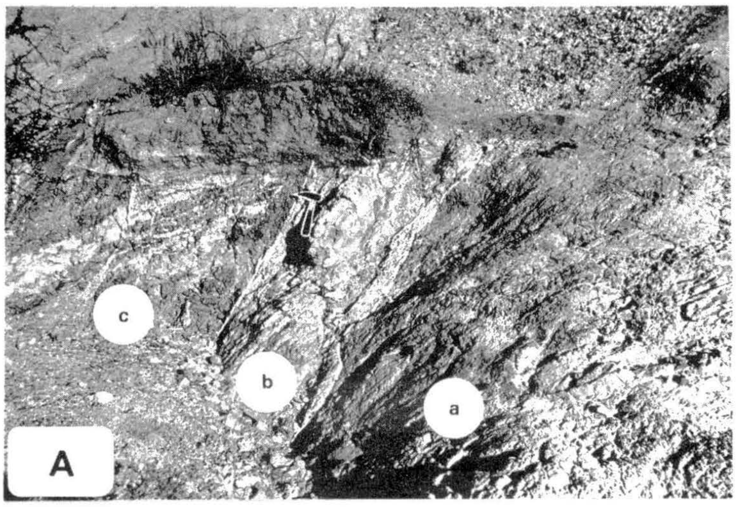
C Welded tuff dyke (light coloured) cutting through the basalt (darker) on Beyers Vley Outspan. (See also Plate I G).

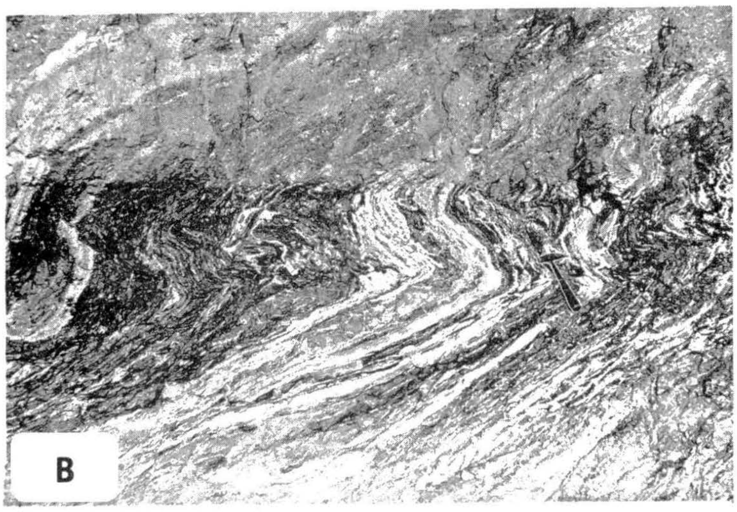
D Vertical view of Enon conglomerate on Riet Fontein, Uitenhage district. Note fractured clasts.

E Enon conglomerate dipping 25° south on Roode Krantz, Uitenhage district. Note imbrication of the quartzite clasts and the irregular sandstone lens.

F Thinly bedded basal Enon sandstone (foreground) overlying basalt with a sharp, straight contact near the Erenkrons Berg - Drie Kuilen boundary, Steytlerville district. Looking south-east.

PLATE 2





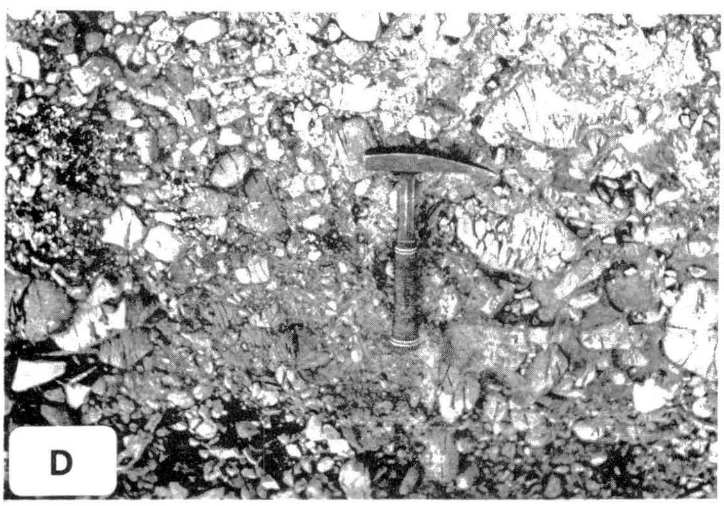


PLATE 3

A Hydrothermally altered tuff (a) with part of an intru= sive dyke: spheroidal weathering dolerite (c) with chill phase (b). Slag Boom, Kirkwood district.

B Enon conglomerate overlying basalt on Slag Boom. The line of contact is remarkably straight. Looking west.

C Enon conglomerate unconformably overlying Witteberg quartzites on Aluin Krantz West, Alexandria district. Looking west.

D Enon conglomerate unconformably overlying Upper Witte= berg sandstones and shales on Haas Poort, Steytlerville district. Looking west.

E Enon conglomerate overlying basalt on Honeyvale, Uiten= hage district. Note straight line of contact. Looking west.

F The western extremity of the Kirkwood Panhandle on Erenkrons Poort, Steytlerville district, looking west. In the foreground the northerly dipping tuff (d) and basalt (e) are overlain by basal Enon sandstones (f) which fill an erosional trough. Farther to the west erosion has stripped off the Mesozoic cover and ex= posed a fossil landscape. This part of the basin is contained within the anticlinal flanks of the Main Witte= berg Quartzite (a) while the Lower Witteberg Shales (b), including the Driekuilen Sandstone (c), form the uneven floor. (See also map and section of Erekroons-poort area).

PLATE 3

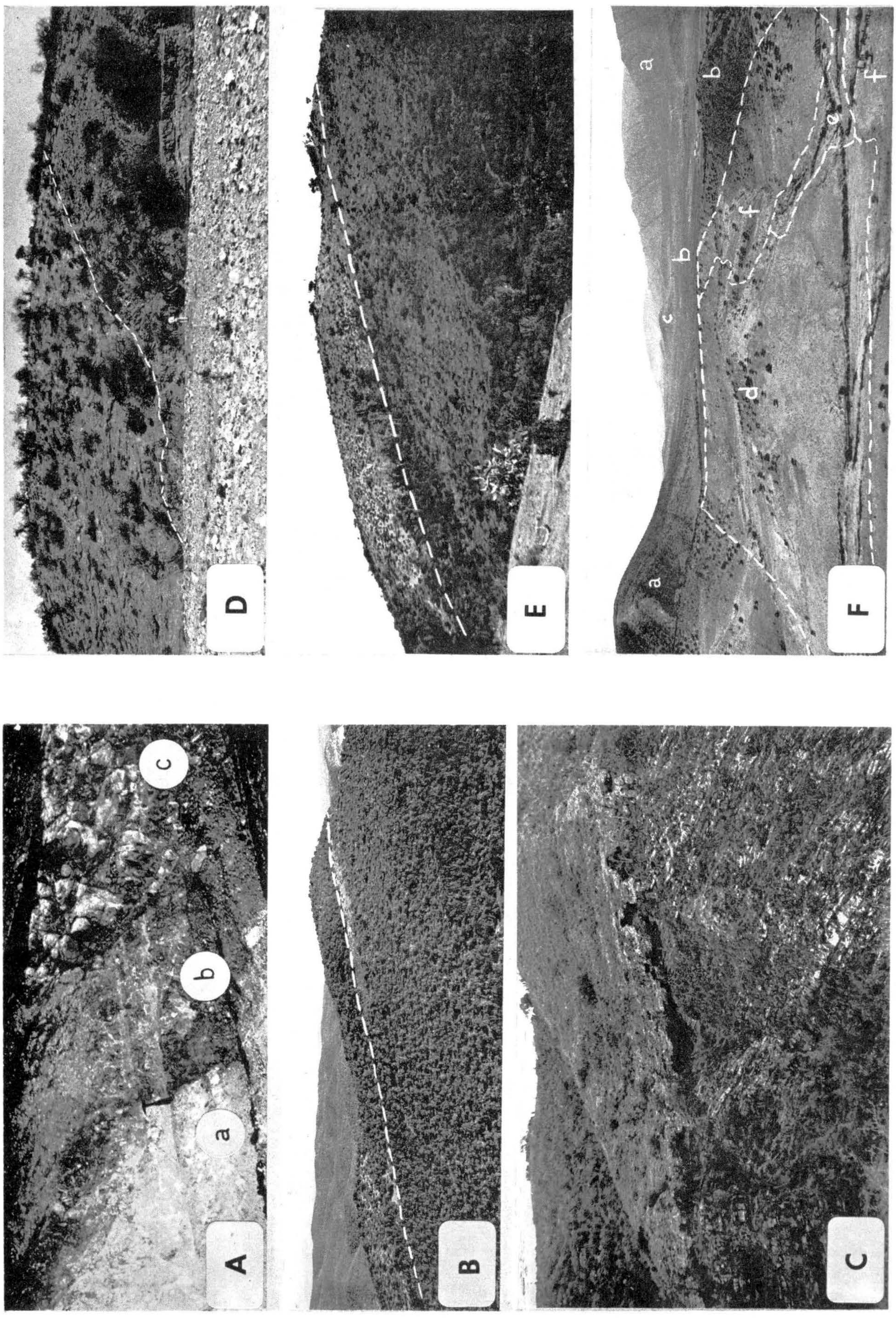


PLATE 4

A Negative print of a pisolitic tuff from Unamore, Alex=andria district. Note concentric structure shown by some pisolites.

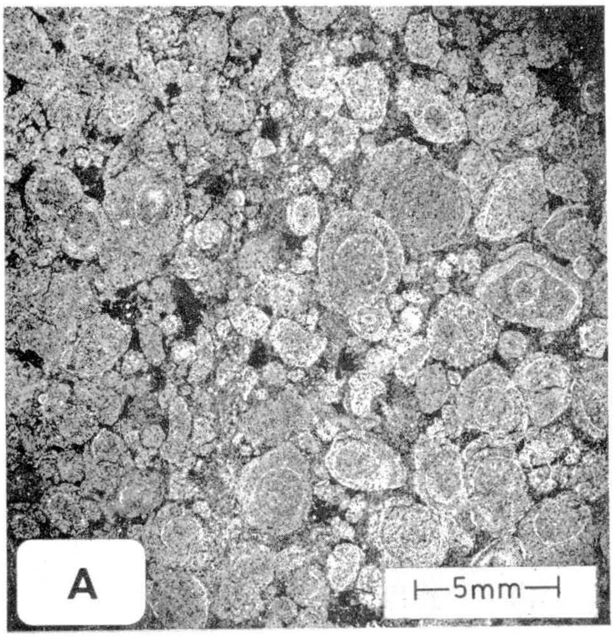
B Photomicrograph of pisolitic (oölitic) tuff from Slag Boom, Kirkwood district. Individual pisolites consist of broken crystals set in very fine volcanic dust. Ordinary light.

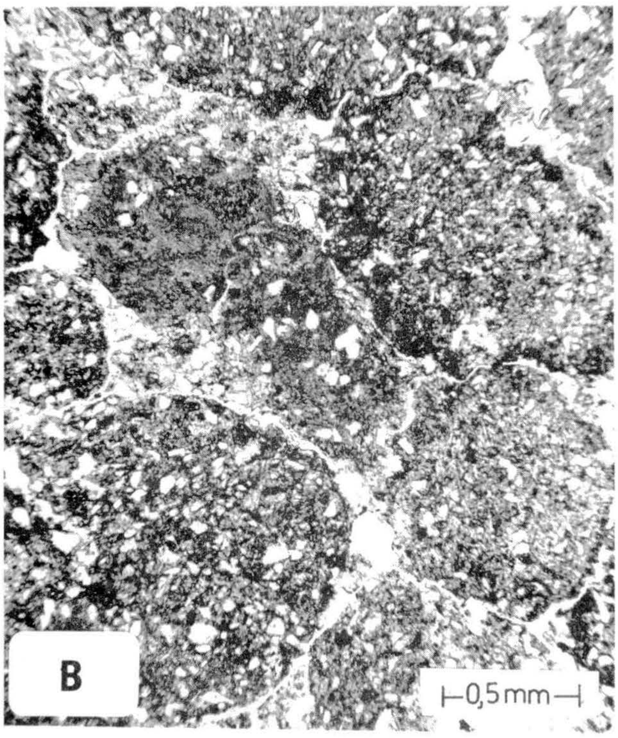
C Vitroclastic texture in a crystal-poor tuff from Beyers Vley Outspan, Steytlerville district. Note various shapes of glass shards. Ordinary light.

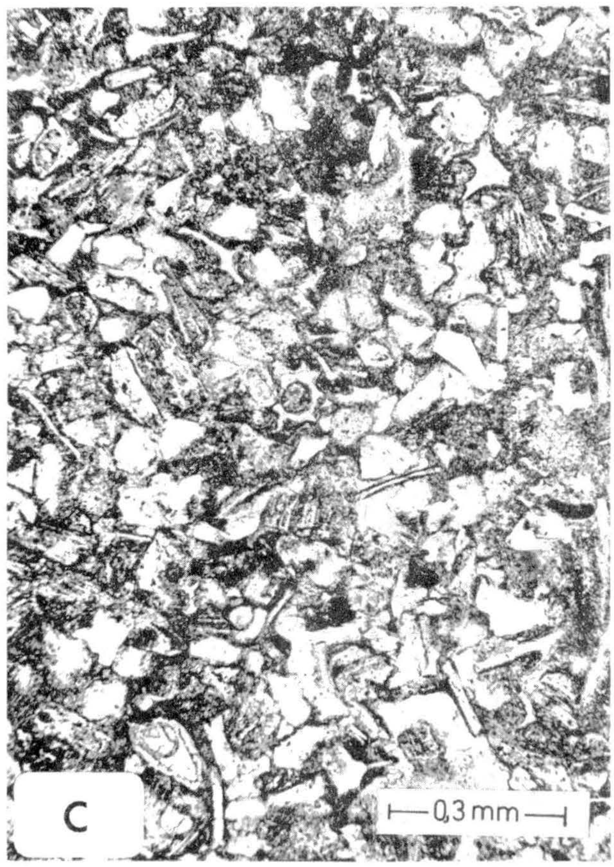
D Vitroclastic texture of a waterlaid tuff from Slag Boom, Kirkwood district. Ordinary light.

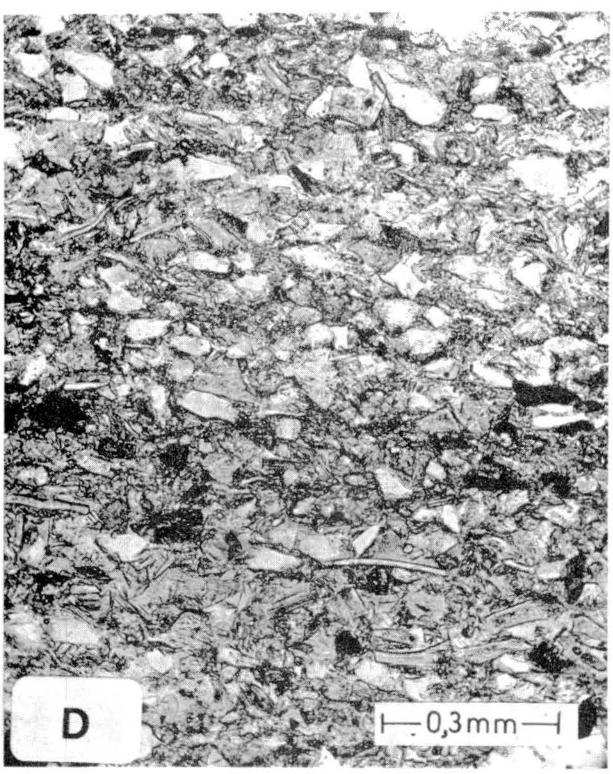
E The same tuff as "D" above with nicols crossed. This tuff has about 23% broken quartz, sanidine, and plagioclase crystals (white and grey). The glass shards remain dark under crossed nicols.

PLATE 4









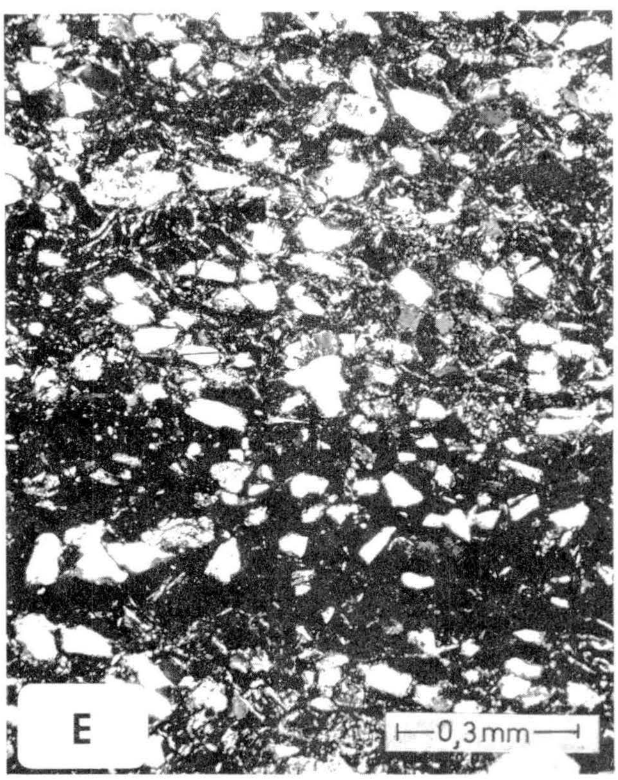


PLATE 5

A Photomicrograph of a typical ash-fall tuff with about 15% broken crystals of quartz, sanidine, and plagio= clase set in a matrix of fine volcanic glass and dust. Ordinary light.

B The same as "A" above under crossed nicols.

C Basal Enon sandstone from Erenkrons Poort, Steytler= ville district. The rock is composed of lithic frag= ments (mottled appearance), quartz and sanidine (grey and white), plagioclase, and glass shards (indicated with arrows). The subopaque matrix consists of silica and iron oxides. Crossed nicols.

D Welded tuff from Slag Boom, Kirkwood district. Some glass shards are bent around phenoclasts. Ordinary light.

E Welded tuff from Enon Mission Station, Kirkwood dis= trict. Note interstitial celadonite (dark grey) and the faint pseudo flow structure in the lower half due to compression and distortion of glass shards. Ordi= nary light.

F Welded tuff from Gorie Laaghte, Uitenhage district, under high magnification. Glass shards are devitri= fied into parallel intergrowths of minerals (usually sanidine and cristobalite), causing an axiolitic structure.

PLATE 5

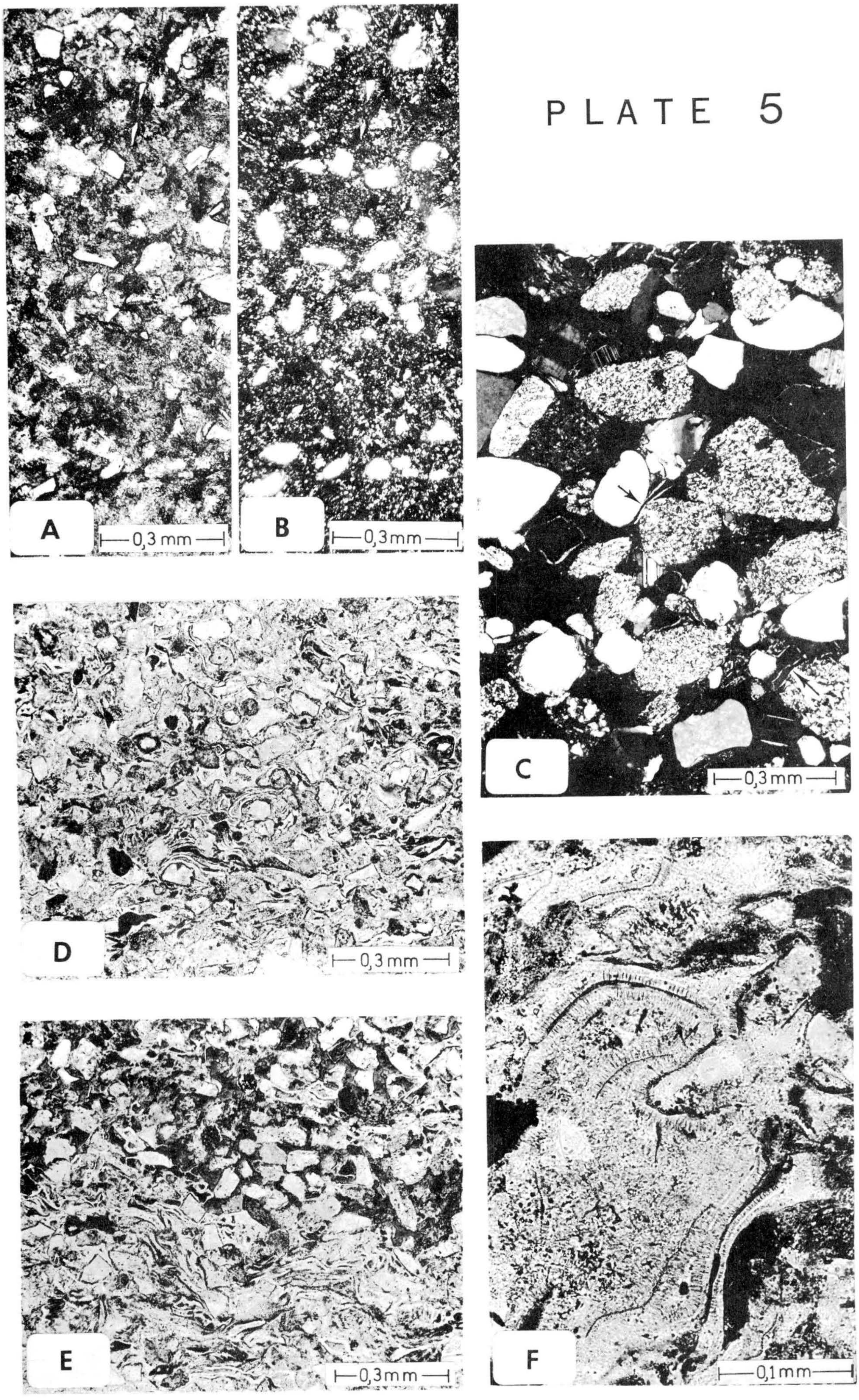


PLATE 6

A Photomicrograph of the contact between the basalt and tuff dyke on Beyers Vley Outspan, Steytlerville district. Note the plagioclase phenocryst protruding into the tuff. Ordinary light.

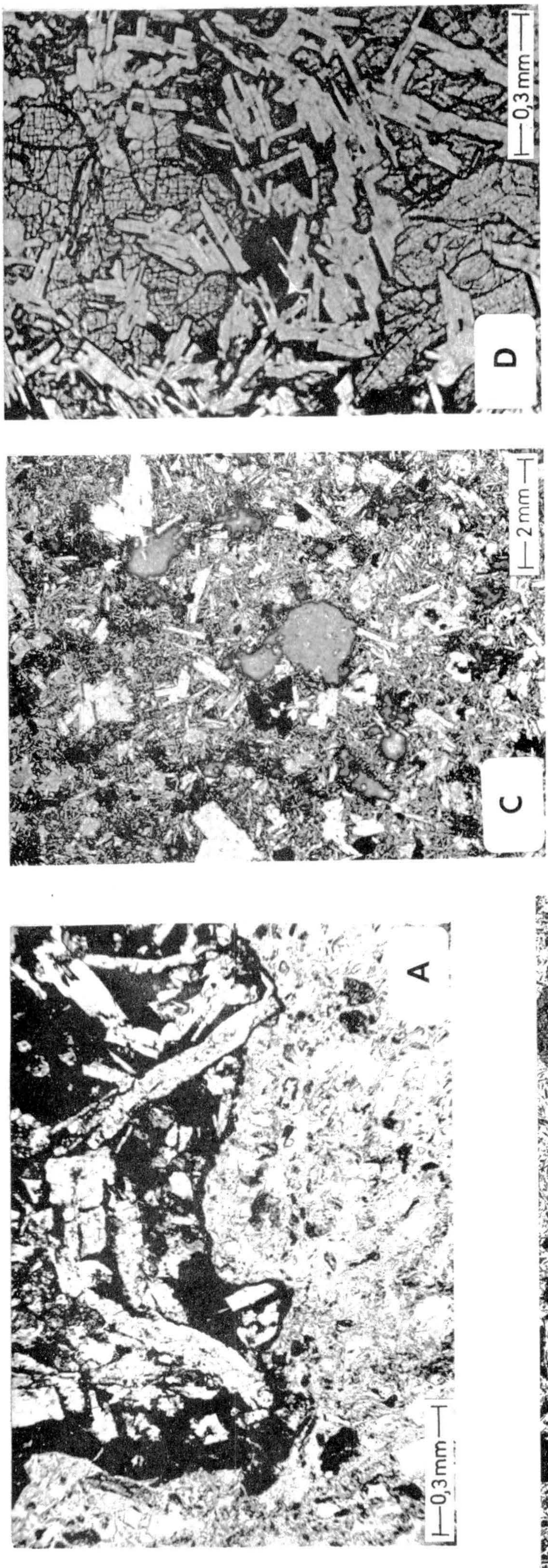
B Negative print of a larger portion of the same contact as in "A" above. The black areas in the basalt are void vesicles. Vesicles in the upper right hand corner are filled with intrusive tuff. The tuff is discoloured along the contact. (See also Plate 1 G).

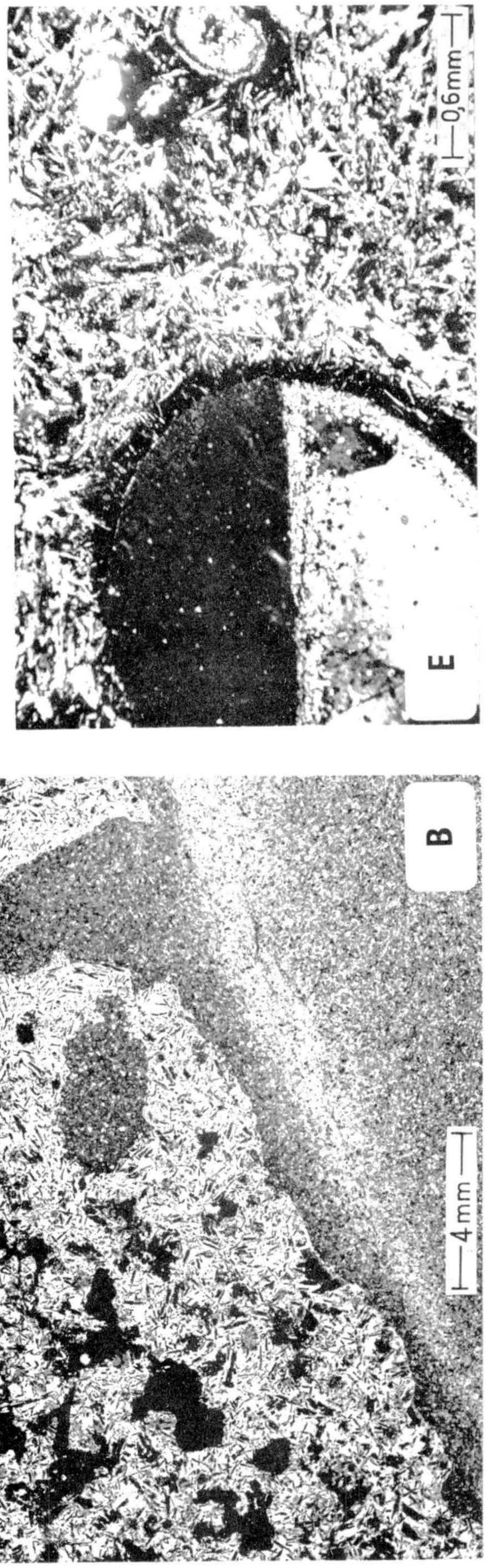
C Amygdaloidal basalt from Gorie Laaghte, Uitenhage district. Iddingsite (black) is pseudomorphous after olivine. Vesicles are lined with celadonite (dark grey) followed by zeolite and calcite (light grey). Ordinary light.

D Dolerite from Slag Boom, Kirkwood district. Plagio=clase and augite are ophitically intergrown. Ordinary light.

E Microcrystalline basalt from Slag Boom with vesicle partly filled with zeolite. The "water level" of the zeolite is parallel to the flow parting of the basalt which now has a dip of about 25° in the field. Crossed nicols.

PLATE 6





APPENDIX

THE TERMINOLOGY AND CLASSIFICATION OF PYROCLASTICS

ABSTRACT

The main objective of a rock classification system is to promote com= munication and it should therefore be simple. Various parameters can be used for classification, but for field use textural properties are probably most important. The development of classifications for pyroclastics over the last few decades is sketched and certain in= consistencies in presently used terminology are pointed out. A simple, flexible classification scheme for pyroclastics is proposed, which endeavours to conform to the usage for other sediments whilst retaining established names as far as possible.

INTRODUCTION

Pettijohn (1957, p. 229-240) gives a clear account of the objec= tives, principles, and problems of rock nomenclature and classifica= tion which is briefly summarized below.

Classification is firstly aimed at promoting communication by provision of appropriate names to classes of related objects. A second objective of classification is the schematic representation of ideas or concepts which make them more comprehensible and easier to remember.

Because it is not feasible to construct a classification system for rocks based on all their properties, only two or three parameters are usually chosen for this purpose. For a workable classification the defining properties should be readily observable, as well as ge= netically significant. A close relation exists between genetic and de= scriptive factors in the classification of sedimentary rocks and even in descriptive classification schemes some degree of commitment as to origin must be made.

In defining classes, either arbitrary "symmetrical" divisions have been chosen or, otherwise, supposed "natural" boundaries have decided the class limits. In many instances they have been dictated by usage or tradition.

Classifications and terminology are inherently subject to adap= tation in the light of new data. But redefinition of terms is to be undertaken with care and then only to sharpen the term rather than radically alter its meaning.

Difficulties are often encountered when sediments of different origin are classified together because those properties significant for one group of sediments may be inappropriate for another. For this reason partial classifications, suitable for a specific category of sediments, have been devised.

Grabau (1913) distinguished between endogenetic (non-clastic) and exogenetic (clastic) rock types. A further distinction between clastic rocks of volcanic (pyroclastic), tectonic (autoclastic), and "normal" (epiclastic) origin is generally made (Pettijohn, 1957, p. 234-235). This simple threefold genetic division of clastic rocks can be used as starting point and each of these categories can be further subdivided according to observable factors.

Except for colour, and to a certain extent structure, the most conspicuous property of a clastic rock in the field is probably its grain size. In modern rock classifications composition is generally ranked first, with texture usually superimposed on composition (Krum= bein and Sloss, 1963, p. 151). For field use however, textural and structural properties of clastic rocks are probably more useful, es= pecially with the finer grained sediments where mineralogical compo= sition cannot be determined adequately in a hand specimen.

Pyroclastic rocks have often been neglected or even excluded from classification systems, probably because of their dual nature. Owing to their endogenetic origin pyroclastics have not been included with "normal" exogenetic sedimentary rocks. Again, their clastic nature "resists" inclusion with igneous rock classifications which are usu= ally restricted to crystalline rocks.

HISTORICAL

Wentworth and Williams (1932) reviewed the classifications and terminology for pyroclastic rocks up to that time and proposed a scheme which was based mainly on size and shape of ejecta, as well as their composition, structure, and origin (whether magmatic or pre-solidi= fied). Table 1 outlines some main points of the classification of Wentworth and Williams (1932, p. 51-53) which has become well esta= blished in the English speaking world, although some authors have in= troduced slight modifications.

Table 1

The classification of Pyroclastic Rocks according to Wentworth & Williams (1932). (Simplified)

Size of particles, mm	Shape of particles	Name of particles	Name of rock
>32	Rounded to subangular, but not water-worn; material partly or entirely molten at time of eruption	Bombs	Agglomerate
	Angular; material solid at time of eruption	Blocks	Volcanic breccia
4-32	Any shape, massive	Lapilli	Lapilli-tuff
	Any shape, vesicular	Cinders	Cindery lapilli-tuff.
$\frac{1}{4}$-4	Any shape	Coarse ash	Coarse tuff
$< \frac{1}{4}$	Any shape	Fine ash or volcanic dust	Fine tuff

Fisher (1961) attempted to group all volcanic clastic rocks together and introduced the term "volcaniclastic". He used the term "volcanic" in a broad sense meaning "volcanically formed" as well as "of volcanic composition" (Fisher, 1963, p. 87). Fisher (1961, p. 1411-1413) subdivided volcaniclastic rocks principally into three types according to manner of production of the fragments:

a) Autoclastic rocks, containing fragments that are produced within (but not usually extruded from) volcanic vents, during movement of lava flows, or by gas explosions within flows that have ceased to flow;

b) Pyroclastic rocks, which consist of fragments produced by volcanic explosion and extruded as discrete particles from volcanic vents;

c) Epiclastic volcanic rocks, containing fragments produced by weathering and erosion of solidified or lithified volca= nic rocks of any type.

In order to include all volcaniclastic rocks within a single system, Fisher found it necessary to define size grades corresponding to those for other clastic rocks. Table 2 embodies Fisher's classification which includes "....a nongenetic category, based only upon particle size and the presence of volcanic material" for rocks with clasts of unknown ori= gin (1961, p. 1409).

Table 2

Proposed Classification of Volcaniclastic Rocks (Fisher, 1961).

Predominant grain size (mm)	Autoclastic *	Pyroclastic * Primary or Reworked		Epiclastic * +	Equivalent non=genetic terms * +
256	Flow breccia, Autobreccia, Intrusion breccia	Pyroclastic breccia, Agglomerate		Epiclastic volcanic breccia,	Volcanic breccia,
64		Lapillistone		Epiclastic volcanic conglomerate	Volcanic conglome= rate
2	Tuffisite	Coarse	Tuff	Epiclastic volcanic sandstone	Volcanic sandstone
1/16		Fine		Epiclastic volcanic siltstone	Volcanic siltstone
1/256				Epiclastic volcanic claystone	Volcanic claystone

* May be mixed with nonvolcanic clastic material

\+ Add adjective "tuffaceous" to rocks containing pyroclastic material < 2 mm in size.

Wright and Bowes (1963) criticized Fisher's use of the term "volca= nic" on the grounds that it had a genetic connotation in the first in= stance and consequently the "epiclastic volcanic rocks", which are not produced by volcanic action, should be excluded from a scheme for "volcaniclastic" rocks. Volcanic breccia was defined by Wright and Bowes (1963, p. 83) as "... a rock composed predominantly of angular frag= ments of any rock greater than 2 mm in size, the brecciation and/or em= placement of which was the result of volcanic action."

The classification of pyroclastics also received considerable atten= tion in the U.S.S.R. and Eastern European countries during the last two decades. Several schemes were proposed of which that of Vlodavets et al. (1963) is probably the most comprehensive. These authors subdivide pyroclastic rocks primarily into two categories:

a) "Extrusives", which consist entirely or predominantly of ex= trusive material, and

b) "Extrusive-sedimentaries", which are either sediments with pyro= clastic components, or redeposited material which is largely or entirely of volcanic origin.

These two main groups are further subdivided into roughly 70 sub-types.

A CRITICAL REVIEW

Any rock classification system should strive at simplification. For clastics of diverse origin, unfounded differences among the para= meters of classification schemes should be avoided.

In this connection Wentworth and Williams (1932, p. 45) wrote: "Consideration of the terminology of the sedimentary rocks must to some degree determine the classification of the pyroclastic, especially in fixing the size-limits between the various types of ejecta, for surely if a more or less uniform set of size-limits or grade-sizes be adopted in subdividing both large groups of rocks, much needless confusion will be avoided."

In the present paper only two factors will be briefly discussed, namely 1) grain size limits, and 2) nomenclature.

1) Grain size limits:

It is rather unfortunate that Wentworth and Williams (1932), while classifying pyroclastics, selected main size classes which differed from those originally proposed for clastic sediments by Wentworth ten years previously.

Although not stated explicitly so, Wentworth and Williams might have considered their proposed size class limits to represent more "natural" boundaries. However, it is to be doubted whether enough is known about pyroclasts to make this assertion. Table 3 which shows various grain size limits that have been proposed for pyroclastics since 1932 rather discredits the possible existence of such "natural" size grade boundaries.

Still other geologists, e.g. Holmes (1965, p. 303) prefer indefinite comparisons of pyroclastic debris to pea, walnut and other sizes.

Ideally, size grade limits should not differ among different cate= gories of clastics. In theory, and probably also in practice, all gradations between different categories exist. This fact in itself calls for standard grain size grades among clastics of different origin.

2) Nomenclature:

This is such a vast field that only some inconsistencies in present usage can be pointed out by way of illustration.

TABLE 3. GRAIN SIZE LIMITS AND TERMINOLOGY FOR UNCONSOLIDATED CLASTIC FRAGMENTS

mm	(EPI-)CLASTIC	P Y R O C L A S T I C								
	Standard usage (Wentworth, 1922)	Wentworth & Williams (1932)	Blyth (1940)	Twenhofel (1950)	Pantó (1959)	Fisher (1961)	Rittmann (1962)		Török (1962)	Vlodavets et al (1963)
256	Boulders	Blocks		Bombs	Blocks	coarse Blocks				Blocks
200	Cobbles								Blocks	
64		&	Blocks	large	&	fine &			&	
50	Pebbles	Bombs			Bombs	Bombs				Bombs
32				Lapilli			Blocks	Bombs	Bombs	
30										
25		Lapilli								
20		&	Lapilli			Lapilli		Cinders		Lapilli
10		Cinders		small-medium				Pumice		
4					Lapilli		Lapilli		Lapilli	Volcanic Gravel
2,5										
2										
1	Sand	coarse	coarse	coarse			volcanic sands			
½			Ash						Volcanic sand	Volcanic sand
¼		Ash		Ash		coarse				
1/10										
1/16			fine		Dust	Ash	Ashes			
1/20	Silt	fine		fine					Volcanic Dust	Volcanic Dust
1/256			Dust			fine	fine			
	Clay									

For instance, the widely used term "lapilli-tuff" is strictly speak= ing not a tuff in the sense defined by Wentworth and Williams. Accor= ding to normal rules of nomenclature "lapilli-tuff" would signify a tuff composed predominantly of lapilli fragments (cf. pebble-conglomerate), in which case the term "tuff", as generally accepted, must be redefined to include particles larger than 4 mm. To avoid confusion it seems desi= rable to replace the term "lapilli-tuff". Fisher (1961) introduced the term "lapillistone", but this is not quite consistent either (compare "pebblestone" with "pebble-conglomerate").

The term "lapilli" (Italian, "little stones") appears to be tho= roughly incorporated into the nomenclature for pyroclastics, although there is some disagreement on whether lapilli fragments should consist predominantly of "essential" (cognate) fragments or whether all ejecta, cognate and noncognate, should be included. The latter definition appears to be more widely recognized (A.G.I. Glossary of Geology, 1960). In this case, should the term "lapilli" be used as the pyroclastic equiva= lent of "pebble" then both "lapilli-agglomerate" and "lapilli-breccia" could claim right of existence in the terminology.

Strictly speaking, a term like "pyroclastic breccia" should also be replaced by one noun with a genetic connotation to eliminate the ad= jective "pyroclastic" (cf. agglomerate vs. conglomerate).

Most tuffs are composed of crystals and crystal fragments set in a much finer matrix of volcanic dust. In the literature these crystals are often being referred to as "phenocrysts" because it is evident that they have crystallized from the parent magma before ejection and fragmen= tation. However, the present author is of the opinion that the term "phenoclast" would be more appropriate because pyroclastic rocks, al= though largely endogenetic, are clastic and not crystalline. The term "phenoclast" was originally used by Pettijohn (1949, p. 30) to describe larger fragments in a non-uniformly sized sediment.

Conclusion: For the sake of simplicity it appears logical to have standard size grade limits for all clastic fragments and rocks, with an equivalent and, preferably, distinctive nomenclature to distinguish between epiclastic, autoclastic, and pyroclastic types.

THE CLASSIFICATION PROPOSED HERE

In the course of an investigation of the Suurberg volcanic rocks the author adopted a classification scheme for pyroclastics which depends primarily on size and degree of rounding of the clasts (Table 4). The endeavour was to conform to the usage for other clastic sediments and to retain established names as far as possible.

Table 4

Proposed Classification for Pyroclastics

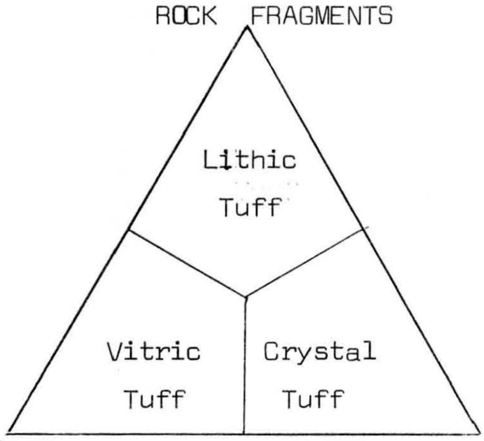
<table>
<tr><th rowspan="2">Size of Clasts (mm)</th><th colspan="3">Rounding & Name of Clasts</th><th colspan="3">Name of Rock</th></tr>
<tr><th></th><th>Angular; solid at time of eruption</th><th>Rounded to sub-angular but not water worn; material partly or entirely molten at time of eruption</th><th></th><th>Angular clasts</th><th>"Rounded" clasts</th></tr>
<tr><td></td><td>large</td><td rowspan="2">BLOCKS</td><td rowspan="2">BOMBS</td><td>coarse</td><td rowspan="3">Pyro=
clastic
BRECCIA</td><td rowspan="3">AGGLOME=
RATE</td></tr>
<tr><td>256</td><td>medium</td><td>medium</td></tr>
<tr><td>64</td><td>small</td><td colspan="2">(Lapilli)</td><td>fine</td></tr>
<tr><td>2</td><td>coarse</td><td colspan="2" rowspan="2">ASH</td><td>coarse</td><td colspan="2" rowspan="2">TUFF</td></tr>
<tr><td>1/16</td><td>fine</td><td>fine</td></tr>
<tr><td>1/256</td><td></td><td colspan="2">(Volcanic Dust)</td><td></td><td colspan="2"></td></tr>
</table>

Using Table 4 as starting point, clasts and rock names may be quali= fied, when necessary, by any of the following factors:-

1) Igneous rock type with which it corresponds chemically (Example: rhyolitic tuff).
2) Composition: Relative proportions of glass shards, crystal fragments, and rock fragments in tuffs (Table 5).
3) Source of constituent pyroclastic particles: <u>Essential</u> — formed directly by crystallization from magma; <u>Accessory</u> — derived from previously solidified volcanic rocks of consangui= neous origin; <u>Accidental</u> — derived from the pre-volcanic basement. (The terms "essential" and "accessory" are more often used in another petrologic sense. Unfortunately they have become widely accepted in pyroclastic terminolgy al= though alternative terms would have been desirable.)
4) Colour.

Table 5

Relative Proportions of Glass Shards, Crystal Fragments, and Rock Fragments in Tuffs (after Pettijohn, 1957)



Note:

Mixtures of pyroclastic material and normal sediments:- When pyroclastic material predominates use nomenclature of normal sediments as modifying adjective (e.g. sandy tuff), and vice versa (e.g. tuffaceous sandstone).

MISCELLANEOUS TERMS:

Unless stated otherwise the following definitions are based, as far as possible, on those of the A.G.I. Glossary of Geology and related Sciences (1960).

Accretionary lapilli (pisolite): Volcanic pellets, commonly exhibiting concentric structures, owing to the accretion of fine ash or dust around raindrops falling through an eruptive cloud, or to accretion around a nucleus fragment which rolls along the ground.

Ash (pyroclastic) fall: Deposition of volcanic ash directly from the air, generally, but not always, resulting in a stratified deposit showing crude to very complete sorting of its component parts (Ross & Smith, 1961, p. 3).

Ash (pyroclastic) flow: A turbulent mixture of gas and pyroclastic materials of high temperature, ejected explosively from a crater or fissure, that travels swiftly down the slopes of a volcano or along the ground surface. The solid material in an ash flow, although unsorted, is dominantly of particles of ash size (Ross & Smith, 1961, p. 3).

Autoclastic: A term applied to rocks that have been brecciated in place by mechanical processes.

(Epi-)clastic: A term applied to mechanically deposited sediments consis= ting of weathered products of pre-existent rocks.

Ignimbrite: A deposit of predominantly welded and unwelded tuff forming extensive and generally thick sheets; probably the result of fiery clouds or pyroclastic (ash) flow.

Pisolitic tuff: An indurated pyroclastic deposit made up chiefly of accretionary lapilli or pisolites.

Pyroclastic: A general term applied to materials that have been explo= sively or aerially ejected from a volcanic vent.

Rework: To move sediment after preliminary deposition.

Tuff (-pebble) conglomerate: A rudite composed mainly of rounded to sub-angular clasts of tuff.

Volcanic: Formed by or pertaining to effusive igneous activity.

Welded tuff: A tuff which has been indurated by heat retained in the particles and the enveloping hot gases. Welding is manifested by fused and deformed glass shards.

CONCLUDING REMARKS

It is realized that the terminology used in the proposed classifi= cation is not entirely satisfactory. For instance, the term "small block" for fragments with a lower size limit of 2 mm seems to be somewhat in= appropriate. As an alternative the term "lapilli" is tentatively sugges= ted but the logical outflow of this has been pointed out (p. vi). The term "pyroclastic breccia", which has been discussed previously (p. vi), is also provisionally included in the proposed scheme. It is not the object of this review to invent new names, but rather to advocate more consistent use of existing terminology.

The foregoing classification scheme does not differ radically from existing ones. The greater simplicity and flexibility introduced should make it useful, particularly in the field.

REFERENCES

AMERICAN GEOLOGICAL INSTITUTE, 1960. Glossary of geology and related sciences. Washington, D.C.

FISHER, R.V. 1961. Proposed classification of volcaniclastic sediments and rocks. Geol. Soc. Amer. Bull., 72, 1409 - 1414.

————————, 1963. Classification of volcanic breccias. Geol. Soc. Amer. Bull., 74, 87.

————————, 1966. Rocks composed of volcanic fragments and their classification. Earth-Sci. Rev., 1, 287-298.

GRABAU, A.W., 1913. Principles of stratigraphy. A.G. Seiler & Co., New York.

HOLMES, A., 1965. Principles of physical geology. Nelson, London.

KRUMBEIN, W.C., and SLOSS, L.L. 1963. Stratigraphy and sedimen= tation. W.H. Freeman & Co., San Francisco.

PANTÓ, G., 1959. Vorschläge zur Schaffung einer einheitlichen Termi= nologie für vulkanische Gesteine. Z. angew. Geol., 5, 373-376.

PETTIJOHN, F.J., 1949. Sedimentary rocks, (1st ed.). Harper & Bros., New York.

————————, 1957. Sedimentary rocks, (2nd ed.). Harper & Bros., New York.

RITTMANN, A., 1962. Volcanoes and their activity. John Wiley & Sons, New York.

ROSS, C.S., and SMITH, R.L., 1961. Ash-flow tuffs: their origin, geologic relations, and identification. U.S. geol. Surv. Profes= sional Paper 366, Washington, D.C.

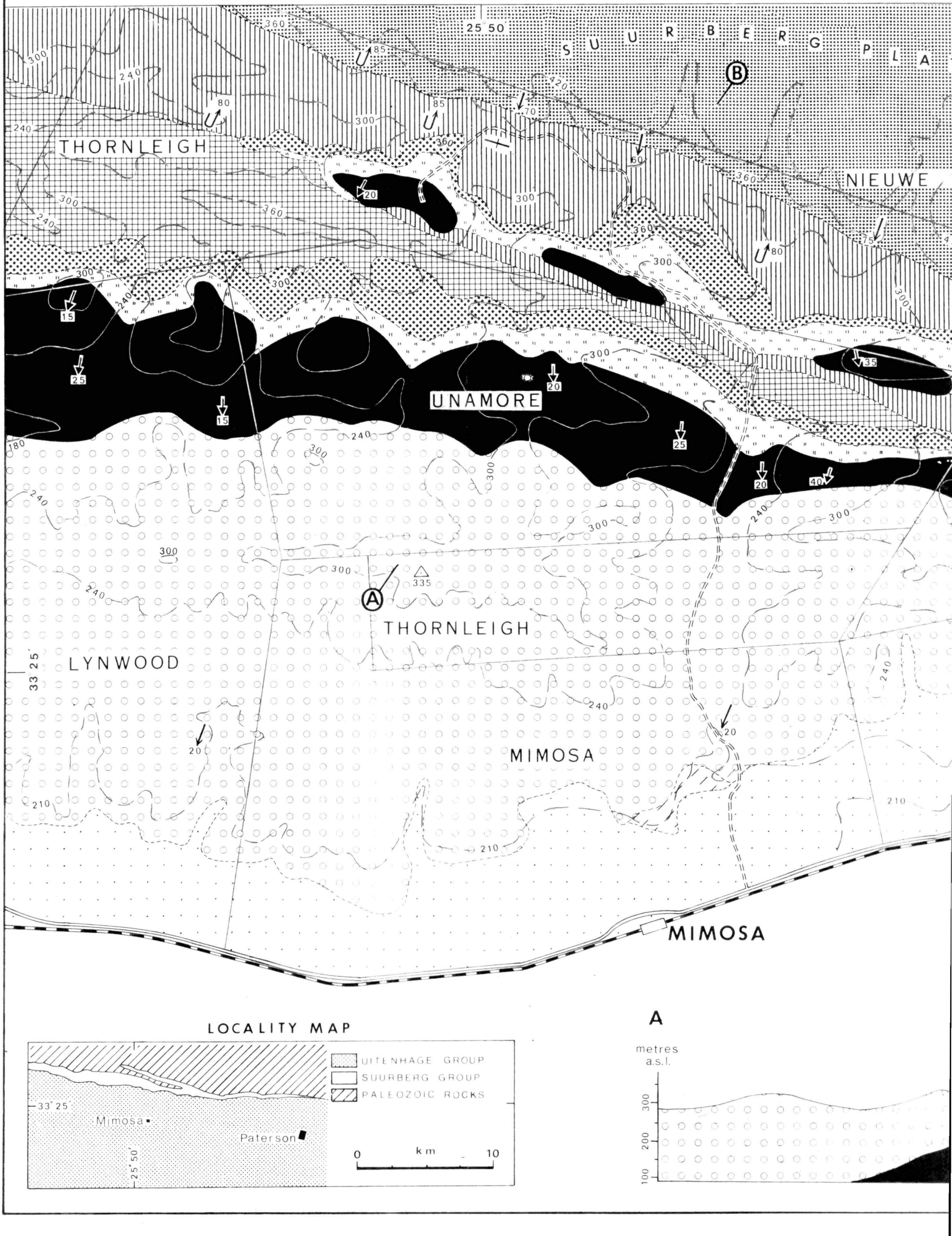
TÖRÖK, Z., 1962. Zur Klassifikation und Studium der pyroklastischen Gesteine. Geologické práce, 25, 183-202.

VLODAVETS, V.I., and others, 1963. A classification of pyroclastic rocks (translated by M.E. Burgunker). Intern. Geol. Rev., 5, 516-524.

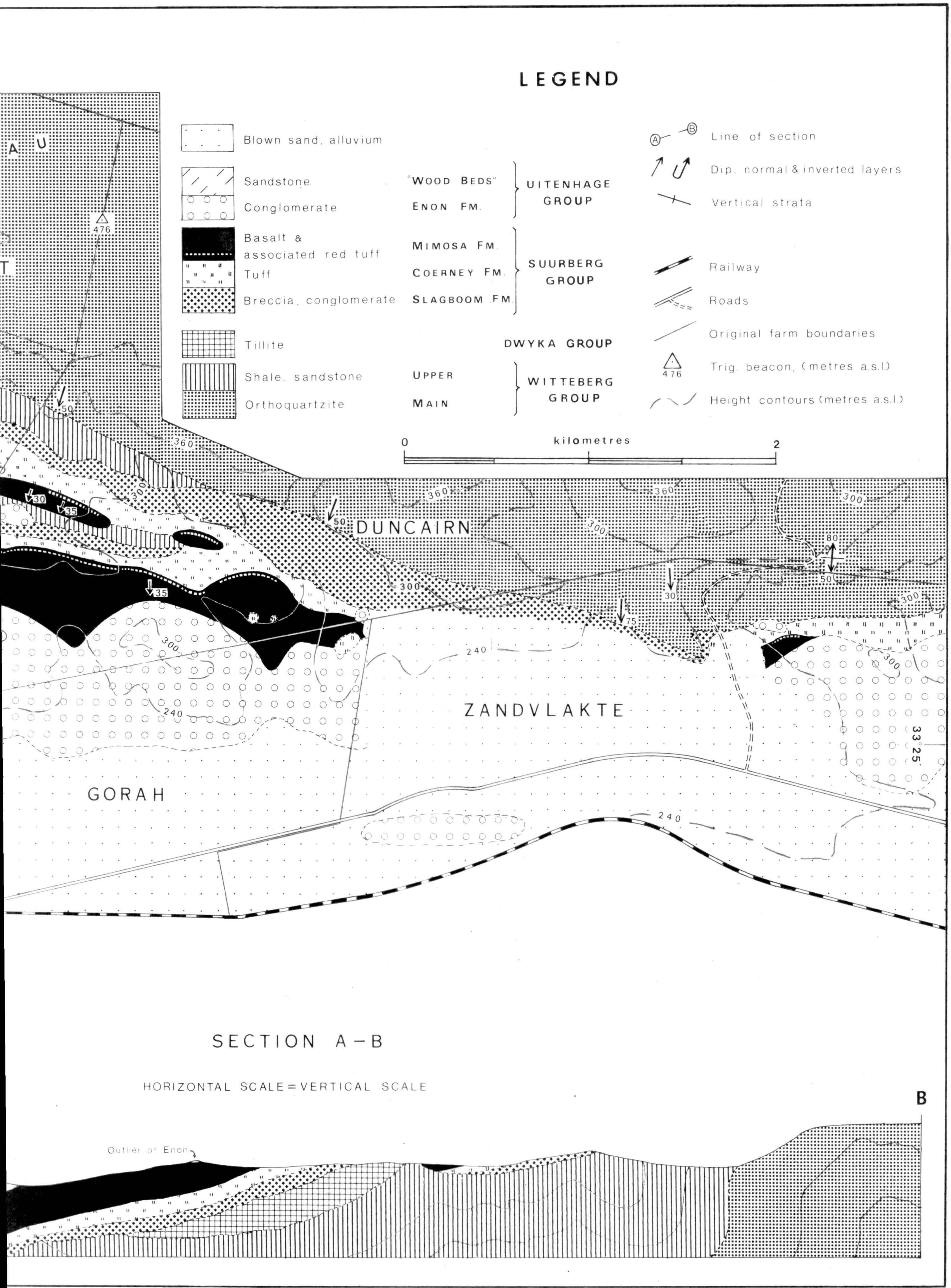
WENTWORTH, C.K., and WILLIAMS, H., 1932. The classification and terminology of the pyroclastic rocks. Nat. Research Council, Rept. Comm. Sedimentation, Bull. 89, 19-53.

WRIGHT, A.E., and BOWES, D.R., 1963. Classification of volcanic breccias. Geol. Soc. Amer. Bull., 74, 79-86.

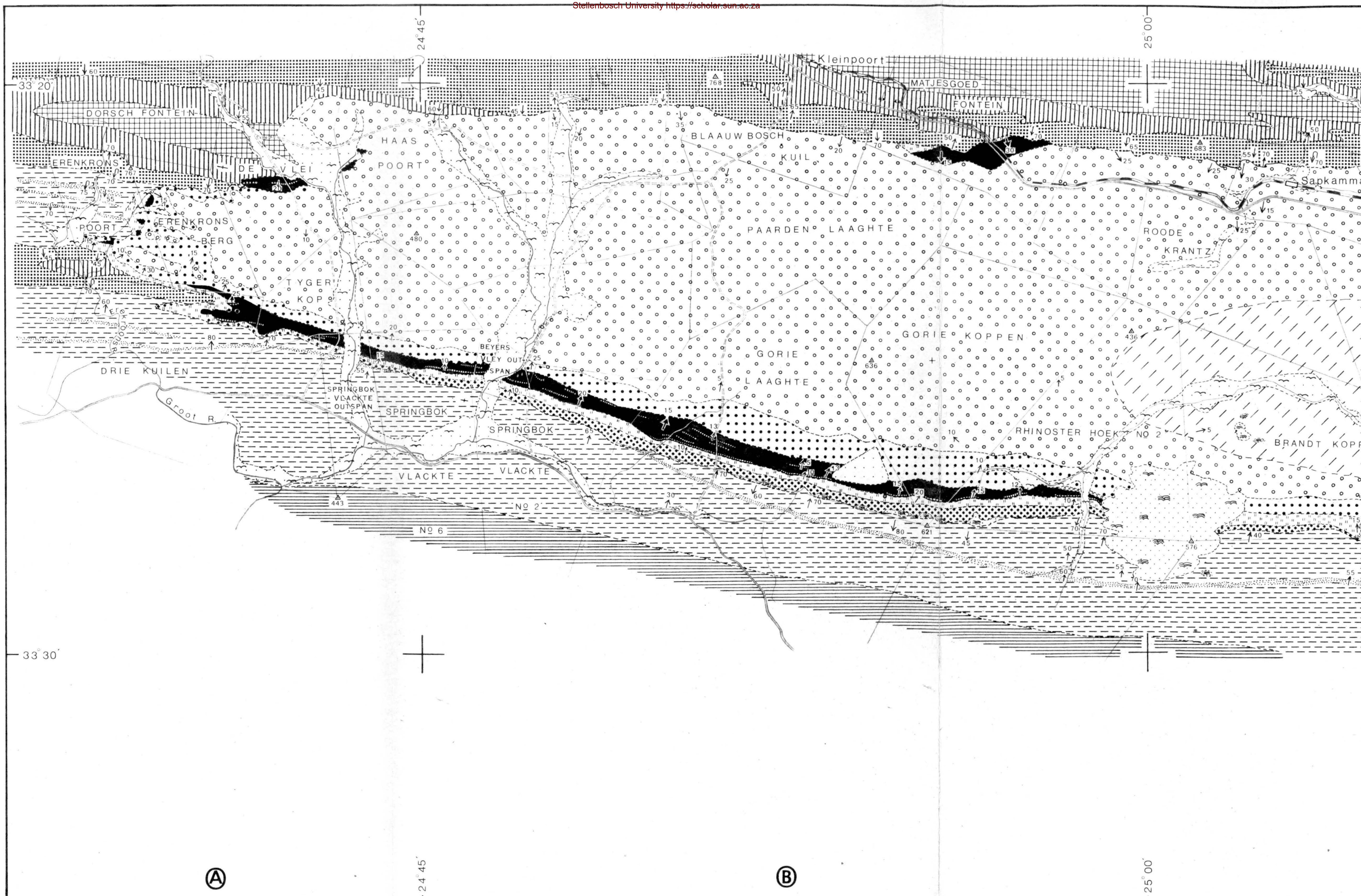
GEOLOGICAL MAP OF THE MIMOSA AREA

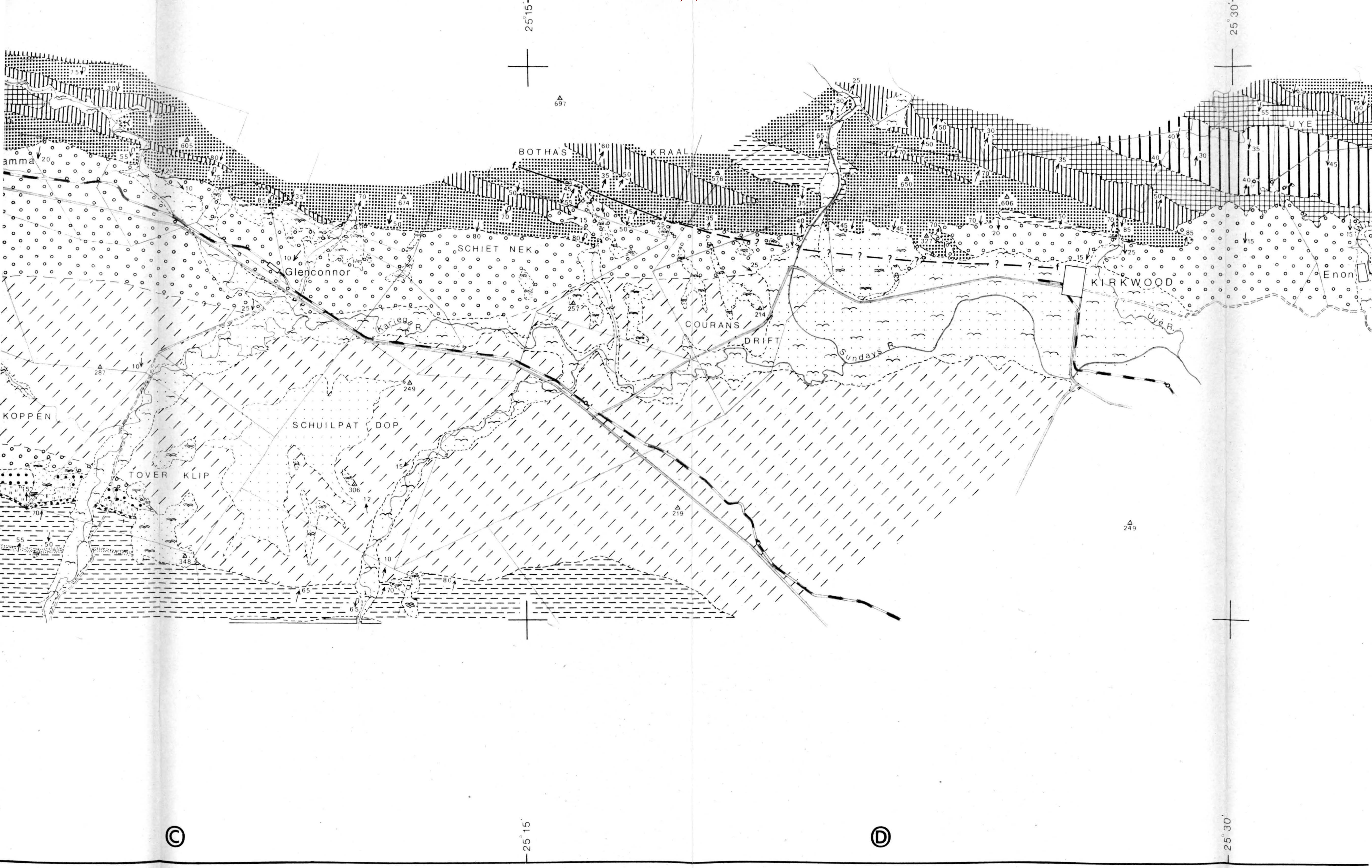


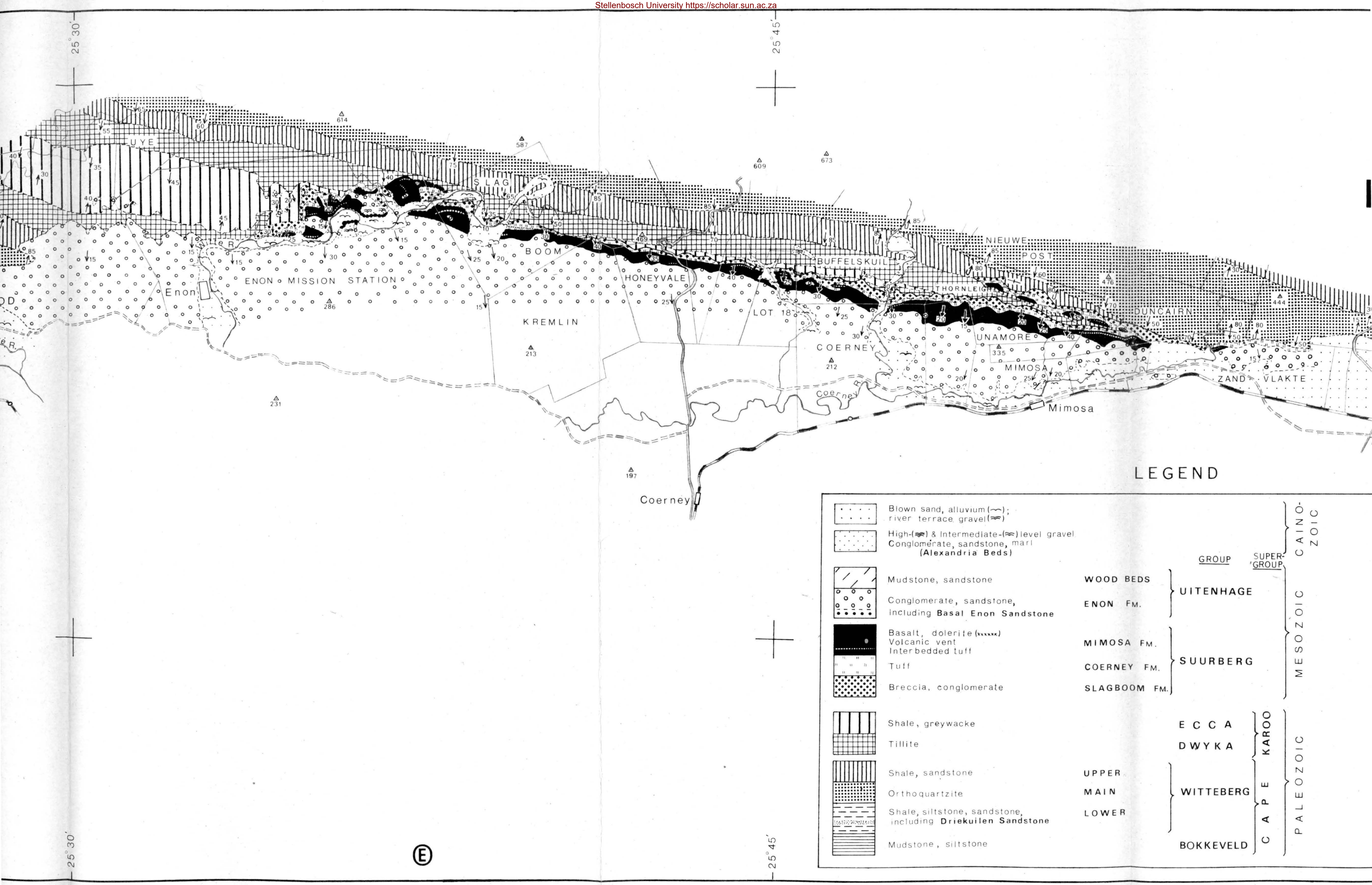
MAP 2

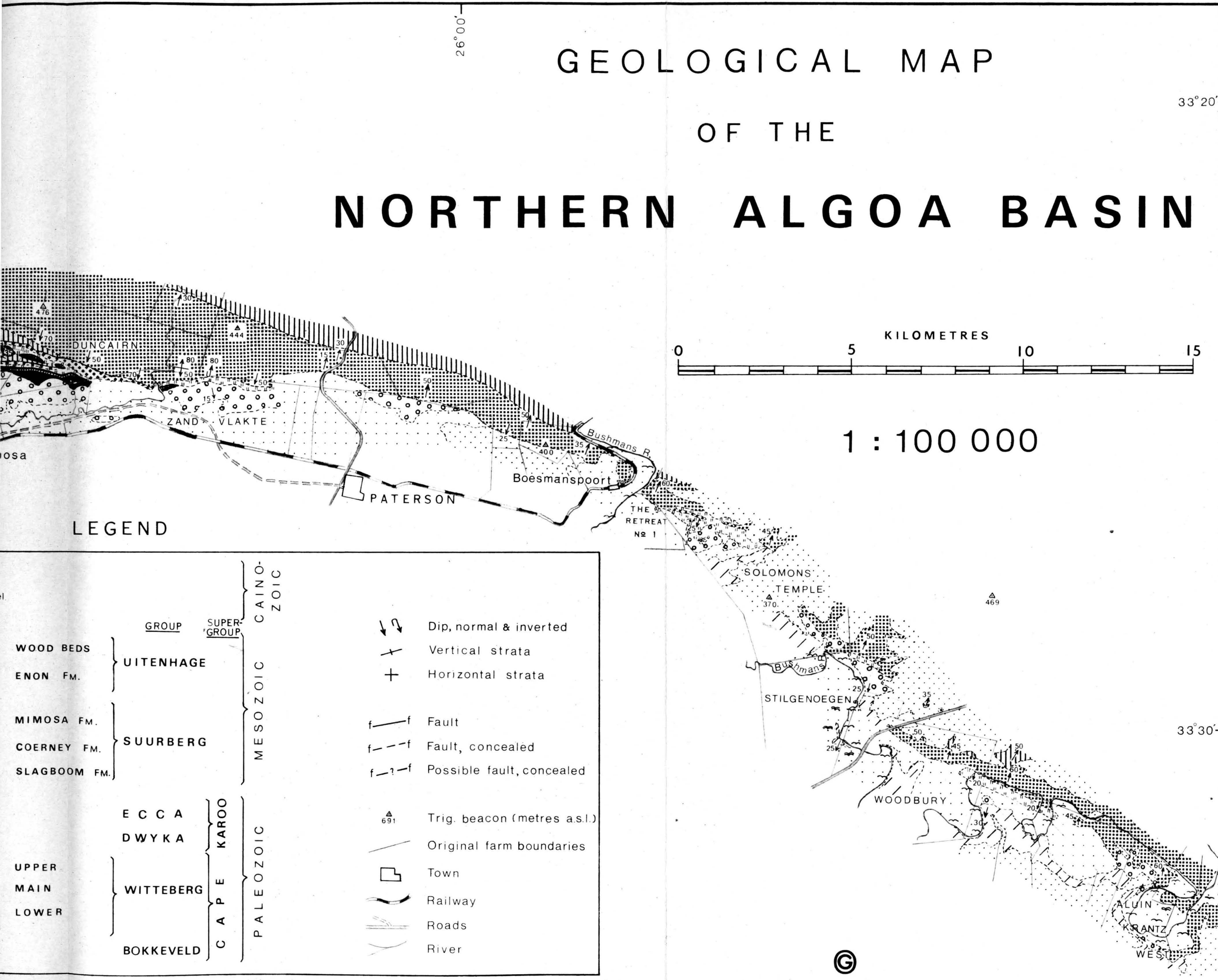

LEGEND
Blown sand, alluvium
Sandstone — "WOOD BEDS"
Conglomerate — ENON FM.
UITENHAGE GROUP
Basalt & associated red tuff — MIMOSA FM.
Tuff — COERNEY FM.
Breccia, conglomerate — SLAGBOOM FM.
SUURBERG GROUP
Tillite — DWYKA GROUP
Shale, sandstone — UPPER
Orthoquartzite — MAIN
WITTEBERG GROUP
Line of section
Dip, normal & inverted layers
Vertical strata
Railway
Roads
Original farm boundaries
Trig. beacon, (metres a.s.l.)
Height contours (metres a.s.l.)
0 kilometres 2
DUNCAIRN
ZANDVLAKTE
GORAH
476
33°25'
SECTION A–B
HORIZONTAL SCALE = VERTICAL SCALE
Outlier of Enon
B

1971


24°45'
25°00'
33°20'
33°30'
Kleinpoort
MATJESGOED
FONTEIN
DORSCH FONTEIN
ERENKRONS
DE VLEI
HAAS
POORT
BLAAUW BOSCH
KUIL
Sapkamma
POORT
ERENKRONS
BERG
PAARDEN LAAGHTE
ROODE
KRANTZ
TYGER
KOP
GORIE KOPPEN
GORIE
LAAGHTE
BEYERS
VLEY OUT-
SPAN
DRIE KUILEN
SPRINGBOK
VLACKTE
OUTSPAN
SPRINGBOK
SPRINGBOK
RHINOSTER HOEK Nº 2
BRANDT KOPPE
Groot R.
VLACKTE
VLACKTE
Nº 2
Nº 6
A
B


25°15'
25°30'
amma
BOTHA'S KRAAL
SCHIET NEK
Glenconnor
KIRKWOOD
Enon
UYE
COURANS DRIFT
Kariega R.
Sundays R.
Uye R.
SCHUILPAT DOP
KOPPEN
TOVER KLIP
697
605
674
576
650
606
327
257
214
287
249
306
219
348
249
C
D
25°15'
25°30'


25° 30'
25° 45'
UYE
SLAG
BOOM
NIEUWE
POST
BUFFELSKUIL
THORNLEIGH
DUNCAIRN
ENON MISSION STATION
Enon
HONEYVALE
KREMLIN
LOT 18
COERNEY
UNAMORE
MIMOSA
ZAND VLAKTE
Mimosa
Coerney
Coerney R.
LEGEND
Blown sand, alluvium; river terrace gravel
High- & Intermediate-level gravel
Conglomerate, sandstone, marl (Alexandria Beds)
CAINO-ZOIC
GROUP
SUPER-GROUP
Mudstone, sandstone
WOOD BEDS
Conglomerate, sandstone, including Basal Enon Sandstone
ENON FM.
UITENHAGE
Basalt, dolerite
Volcanic vent
Interbedded tuff
MIMOSA FM.
Tuff
COERNEY FM.
Breccia, conglomerate
SLAGBOOM FM.
SUURBERG
MESOZOIC
Shale, greywacke
ECCA
Tillite
DWYKA
KAROO
Shale, sandstone
UPPER
Orthoquartzite
MAIN
Shale, siltstone, sandstone, including Driekuilen Sandstone
LOWER
WITTEBERG
Mudstone, siltstone
BOKKEVELD
CAPE
PALEOZOIC
E


GEOLOGICAL MAP
OF THE
NORTHERN ALGOA BASIN
26°00'
33°20'
33°30'
KILOMETRES
0
5
10
15
1 : 100 000
DUNCAIRN
ZAND VLAKTE
PATERSON
Boesmanspoort
Bushmans R.
THE RETREAT № 1
SOLOMONS TEMPLE
STILGENOEGEN
WOODBURY
ALUIN KRANTZ WEST
LEGEND
GROUP
SUPER-GROUP
CAINO-ZOIC
WOOD BEDS
ENON FM.
UITENHAGE
MIMOSA FM.
COERNEY FM.
SLAGBOOM FM.
SUURBERG
MESOZOIC
ECCA
DWYKA
KAROO
UPPER
MAIN
LOWER
WITTEBERG
BOKKEVELD
CAPE
PALEOZOIC
Dip, normal & inverted
Vertical strata
Horizontal strata
Fault
Fault, concealed
Possible fault, concealed
Trig. beacon (metres a.s.l.)
Original farm boundaries
Town
Railway
Roads
River

1971